A PALYNOLOGICAL INVESTIGATION INTO PALAEOENVIRONMENTAL CHANGES IN THE EARLY CENOZOIC SEDIMENTS OF SEYMOUR ISLAND, ANTARCTICA

Jennifer C. Greenhalgh

University College London

Ph.D. Geological Sciences

Abstract: A palynological investigation into palaeoenvironmental changes in the Early Cenozoic sediments of Seymour Island, Antarctica.

The Paleocene and Eocene sediments outcropping on Seymour Island, Antarctic Peninsula have been investigated for their dinoflagellate cyst and sporomorph content. The data have been quantitatively analysed to provide a high-resolution study of palaeoenvironmental change with an emphasis on integrating the marine and continental records, as well as to update the biostratigraphical potential of the palynomorphs. Samples were collected from 10 sites across Seymour Island and the specimens were analysed to produce absolute abundance data and ratio curves of sporomorph/dinoflagellate cysts, Peridinioid/Gonyaulacoid dinoflagellate cyst and species richness and absolute abundance. The count data were also analysed statistically by Correspondence Analysis to provide further details of species abundance and composition within assemblages in order to highlight palaeoenvironmental change and its impact. The evidence suggests widespread dramatic changes in marine conditions in the earliest Danian, before more stable conditions developed which may be related to climate warming in the Late Paleocene. Marine conditions also varied widely in the Eocene, but these appear to have been more localised changes relating to regional tectonic activity, while the climate seems to have undergone small-scale warming and cooling events. Evidence exists for the existence of a cooling trend in the lowermost Paleocene. Following a mid-Paleocene sedimentary hiatus, both marine and terrestrial palynomorph data suggest the location of the Eocene Climatic Optimum within the lower La Meseta Formation. Climatic

cooling, presumably linked to the development of the cryosphere as the continent became isolated, is also observed.

Acknowledgements

I would like to thank the following for the part that they have played in enabling me to finish and submit this thesis. Firstly, the Natural Environment Research Council, the British Antarctic Survey and the Royal Liverpool Seamans' Widows and Orphans Fund, for funding my doctoral studies. Secondly, my supervisors: Professor Alan Lord of University College London, Jamie Powell of Dinosystems and Dave Cantrill, formerly of the British Antarctic Survey, now of the Swedish Museum of Natural History.

I would also very much like to thank the staff of the University College London Geological Department and British Antarctic Survey for their support and encouragement. In particular, Jim Davy of the UCL Geological Department Technical Staff, for all his efforts in helping me over the duration of this project, and for proving that you can get the staff these days; Pete Rawson, formerly Postgraduate Tutor, for his continued enthusiasm and discussions, and Sandra Nederbragt, also of the UCL Geological Department for her help with the statistical analysis carried out. I would also like to thank Dr John Williams of the Natural History Museum, London, for the use of his Palynology Library.

On a more personal note, I would like to thank the following: my mother, for being so supportive and generally understanding during my years of study, illness, and more study; Helen Swan, for her continued support and friendship, and Ben Pluck, for helping me through the writing-up. I would also like to thank those other postgraduate students that I have worked alongside for the past four years, for their friendship, company, and for providing proof that it is possible to finish and live to tell the tale.

Table of Contents

Table of Figures

Chapter 1 - GENERAL INTRODUCTION

1.1 - BACKGROUND AND AIMS OF THESIS

Throughout the Mesozoic, Antarctica was part of the southern supercontinent of Gondwana. The pattern of land distribution at the time caused equatorial oceanic currents to carry warm water to the high latitudes, ameliorating the climate and ensuring that the poles were free of permanent ice (Berggren and Hollister, 1977). The present-day continent of Antarctica is isolated at the South Pole, Gondwana having rifted into Africa, Antarctica (East and West), South America, India, Madagascar, Australia and New Zealand in the Cretaceous and Early Cenozoic. This progressive geographic isolation led to the development of the Antarctic Circumpolar Current (Lawver and Gahagan, 1998). The current, which also developed progressively, prevents warm water reaching Antarctica from the north. However, permanent ice may have developed as early as the Late Eocene in East Antarctica, although grounded ice sheets did not develop throughout the entire region until it had been completely isolated (Cape Roberts Science Team, 2000).

Prior to the isolation of Antarctica, there had been a diverse terrestrial and marine flora (Cranwell, 1959) and fauna (Rich *et al.* 1989, Molnar, 1989). As Antarctica became increasingly isolated, the flora became correspondingly specialised as their surroundings changed and showed increasing endemism (Askin, 1989). As the climate deteriorated further, and permanent ice began to accumulate, conditions for life became increasingly difficult, until by the Miocene the development of grounded ice sheets caused the extinction of most forms of plant life on the continent (Dingle & Lavelle, 1998). The floras of Antarctica, both terrestrial

and marine, recorded the details of this change in their patterns of abundance and distribution. These can be detected today both by the presence of the parent plant body fossils and fossilised spores and pollen (in the case of the terrestrial flora), and the fossilised resting cysts of dinoflagellates (in the case of the marine flora).

The aim of this thesis is to use the terrestrial and marine palynofloras from Paleocene and Eocene sediments that crop out on Seymour Island, on the Antarctic Peninsula, to achieve a high resolution analysis of palaeoenvironmental changes that occurred through the Paleogene, using the palynofloras of this key area as a proxy. The high latitude oxygen isotope record shows progressive cooling through the Eocene from a peak of warmth in the Late Paleocene and Early Eocene (Shackleton & Kennett, 1975). These high latitude areas are also more susceptible to climate changes and this is where the effect is most marked. The intention is to study temperature and environmental fluctuations, looking at the reaction of the palynofloras as temperatures increase and decrease in order to build up a more detailed illustration of the fine-scale changes in regional climate change.

1.2 - GEOLOGICAL HISTORY

1.2.1 - Geological Evolution of Antarctica

Gondwana, the southern landmass of the supercontinent Pangaea, was formed during the Late Palaeozoic. It comprised the modern landmasses of South America, Africa, Australia, New Zealand, Antarctica (East and West), India and Madagascar when it began to break up in the Mesozoic. Until this time, East Antarctica lay in a central position in the land mass, adjacent to all areas that would become separate continents.

The precise mechanism of the break-up of Gondwana is still the subject of debate (Cox, 1992) as the relationship between the roles played by mantle plumes, continental flood basalts, subduction and extension is unclear. The earliest evidence for substantial geological activity in Gondwana is major vulcanism at 182 ± 2 Ma, linked to the Karoo plume that was centred under southeast Africa (Hooper *et al.* 1993) (Fig. 1.1A). Lavas and igneous intrusions that occur throughout East Antarctica and southeast Africa today represent this within-plate magmatism.

While this was volcanism was occurring, West Antarctica, which comprises a number of smaller crustal blocks (Antarctic Peninsula, Thurston Island, Marie Byrd Land, Ellsworth-Whitmore Mountains and the Haag Nunataks) and southern South America, had formed a magmatic arc along the proto-Pacific margin of southwest Gondwana (Storey *et al.*, 1992). The magmatic arc formed as a consequence of the subduction of the proto-Pacific plate underneath an active extensional margin, beginning in the Triassic and continuing throughout the Jurassic (Pankhurst, 1982). Into the fore- and back-arc basins, formed in this extensional area, a large amount of sediment was deposited. A further result of subduction along this margin was the development of an extensional province in southern South America and West Antarctica to the east of the magmatic arc, in the Weddell Sea at 160-156 Ma, (LaBreque *et al.* 1986).

Simultaneously with the subduction occurring on the southwest coast of Gondwana, subduction was also occurring on the northeast margin of Gondwana as the Palaeo-Tethys plate was subducting under Gondwana. This would have added further to the extensional east-west stresses on Gondwana as a whole (Storey *et al.*, 1992).

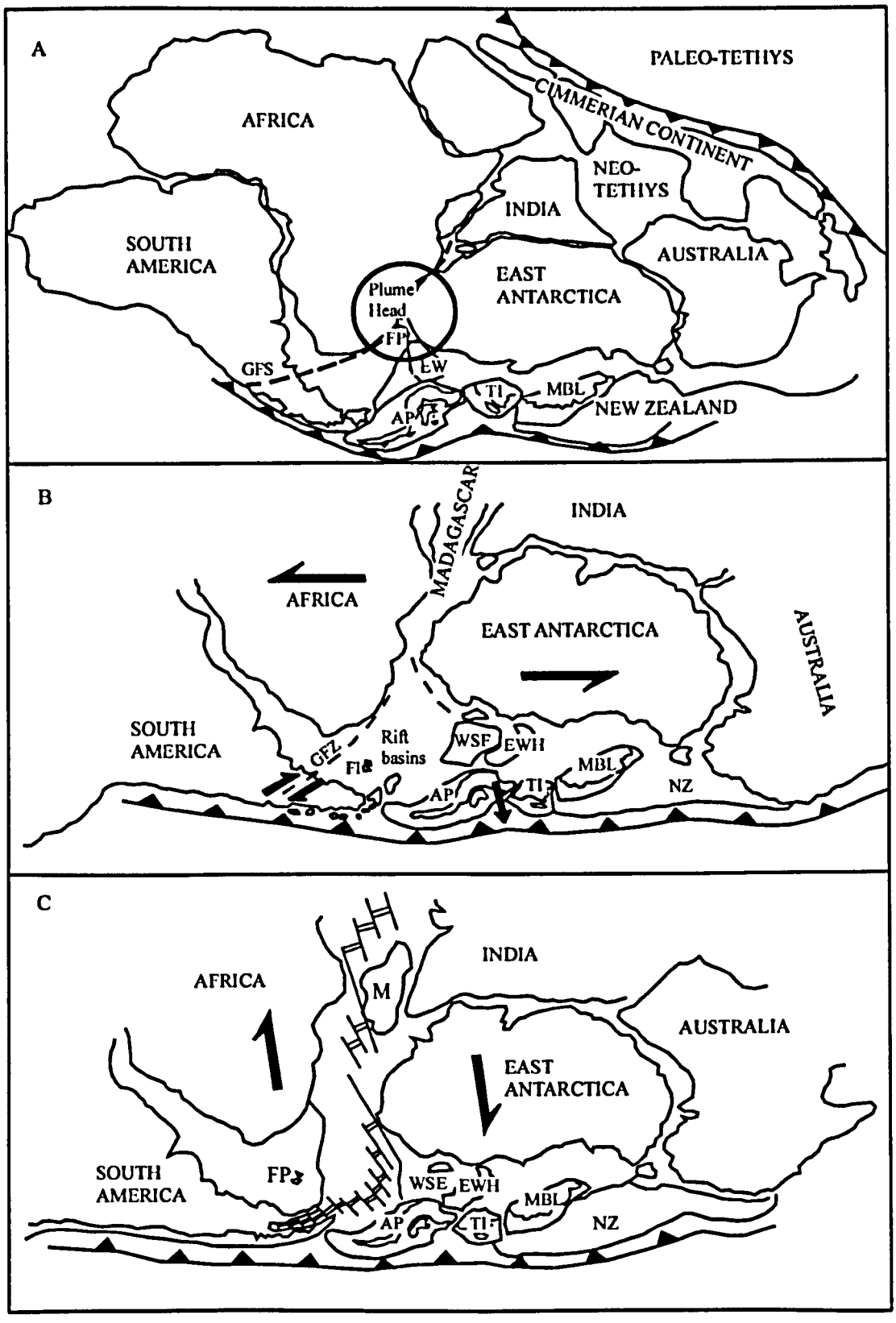


Fig. 1.1. Breakup of Gondwana, after Storey *et al.* (1992, 1996). A, Early Jurassic reconstruction of Gondwana, with position of Karoo plume head and subduction zones. B, Late Jurassic rifting between East and West Gondwana, and rifting in the Weddell Sea. C, Early Jurassic North-South drift between East and West Gondwana. AP, Antarctic Peninsula; EWH, Ellsworth-Whitmore Mountains & Haag Nunatak block; FP, Falklands Plateau; GFZ, Gastre fault zone; M, Madagascar; MBL, Marie Byrd Land; NZ, New Zealand; TI, Thurston Island; WSE, Weddell Sea Embayment.

The separation of East and West Gondwana began pre-157 Ma, based on the age of the M22 anomaly of the earliest formed seafloor in the Somali Basin (Simpson *et al.*, 1979). The East Antarctica continent moved in a northeast direction forming the Somali Basin between the two blocks before movement ceased. This motion may have been related to reactivation of basement shear zones in central and southern Africa (Daly *et al.*, 1989, 1991), amongst which was the Limpopo belt. According to Cox (1992) the Limpopo area developed into a sinistral shear zone along which East Antarctica moved. Alternatively, Storey *et al.* (1992, 1996) related this movement to activity along the southern South American Gastre Fault System which links to the Aghulas Fracture Ridge (Rapela & Pankhurst, 1992) in southern Africa (Fig. 1.1B).

A further stage in the break up of Gondwana, and the evolution of Antarctica, occurred as both Madagascar and East Antarctica began to move south, relative to Africa. This was due to the development of a spreading ridge and associated north-south transform faults (Cox, 1992) (Fig. 1.1C). Storey *et al.* (1996) date this change in the direction of plate motion to 150-140 Ma.

South America began to break away from Africa as the South Atlantic began to open initially in the Late Jurassic and then more significantly during the Early Cretaceous (Wilson, 1992). Associated with this event was the development of two major plumes, the St Helena and Tristan do Cunha, both still extant, and associated with flood basalts in both South America and Africa (Wilson, 1992). The two continents began to diverge along the mid-Atlantic ridge, and this east-west drift combined with the existing southwards movement of east Antarctica to form a three plate drift configuration.

India began to separate from Antarctica during the Early Cretaceous at 135 Ma (Roeser, 1996), while the further movements of Africa, Antarctica and South

America caused the formation not only of the early South Atlantic, but also of the Indian Ocean. New Zealand separated from West Antarctica during the Late Cretaceous at 84 Ma (Storey *et al.*, 1999). Australia separated from Antarctica at about 35.5 Ma (Storey *et al.*, 1999), and South America separated from Antarctica during the Miocene (Barker & Burrell, 1977).

The effect of the break-up of Gondwana on the continent of East Antarctica and the smaller crustal blocks that make up West Antarctica, was to leave the formerly central landmass isolated at the south pole. It had earlier been described by Berggren & Hollister (1977), as being subject to the influence of equable climates and oceanic thermohaline homogeneity. By the end of the Paleogene, its thermal isolation at the pole led to high latitude cooling and the development of permanent ice cover (Zachos *et al.*, 1994).

Both East and West Antarctica, although currently adjacent to each other, remain separate entities with different origins and histories. East Antarctica is a single landmass, while West Antarctica comprises several smaller microcontinents separated from each other, and East Antarctica by past seaways. Their glaciation history is also thought to differ: East Antarctica is thought to have developed permanent ice several million years earlier than West Antarctica (Cape Roberts Science Team, 2000). This could possibly be due to the development of a circum-East Antarctic current moving around East Antarctica and through the seaway then existing between the Weddell and Ross Seas. This could have cooled East Antarctica prior to the formation of the Circum-Antarctic Current through the Drake Passage (Cape Roberts Science Team, 2000) as South America moved north away from the Antarctic Peninsula at around the time of the Eocene-Oligocene boundary. The passage was fully open by 30 Ma (Lawver and Gahagan, 1998).

1.2.2 - Geological Evolution of the Antarctic Peninsula

The Antarctic Peninsula is one of the crustal blocks, or microcontinents, that make up West Antarctica. It has previously been described as being formed as an arc-trench system (Storey & Garrett, 1985) during the Mesozoic (the basement itself having been formed earlier) as a consequence of the subduction of the proto-Pacific plate along the southwestern coast of Gondwana (Barker 1982). More recent reconstructions have identified the Peninsula as representing a terrane-continent collision zone (Vaughan & Storey, 2000). At least three terranes of parautochthonous or allochthonous origin are proposed to be represented (Vaughan & Storey, 2000).

The Antarctic Peninsula was subject to a prolonged period of extension throughout the Late Triassic to Late Jurassic (Storey *et al.*, 1996). During this time, several pulses of magmatic activity occurred (Pankhurst, 1982), some of which are recorded as dykes and sills in the sediments deposited in back-arc basins, such as the Latady and Larsen Basins (Storey *et al.*, 1996). Both of these basins lay on the eastern margin of the arc (Fig. 1.2), and contain sediments deposited under marine conditions e.g. the Nordenskjöld Formation (Farquharson, 1982) (see 1.3.1 below).

During the Early Cretaceous, extension continued, further amounts of magma were emplaced, and large amounts of sediment were deposited in back-arc basins (Storey *et al.*, 1996). The sediments began to record a wider variety of environments, with evidence of turbidite sequences in the southern arc area (Storey et al., 1996). Further north on the arc, the Larsen Basin has contemporaneous deposits that are submarine fan conglomerates (Thomson & Farquharson, 1984), both of which are indicative of deep water sedimentation.

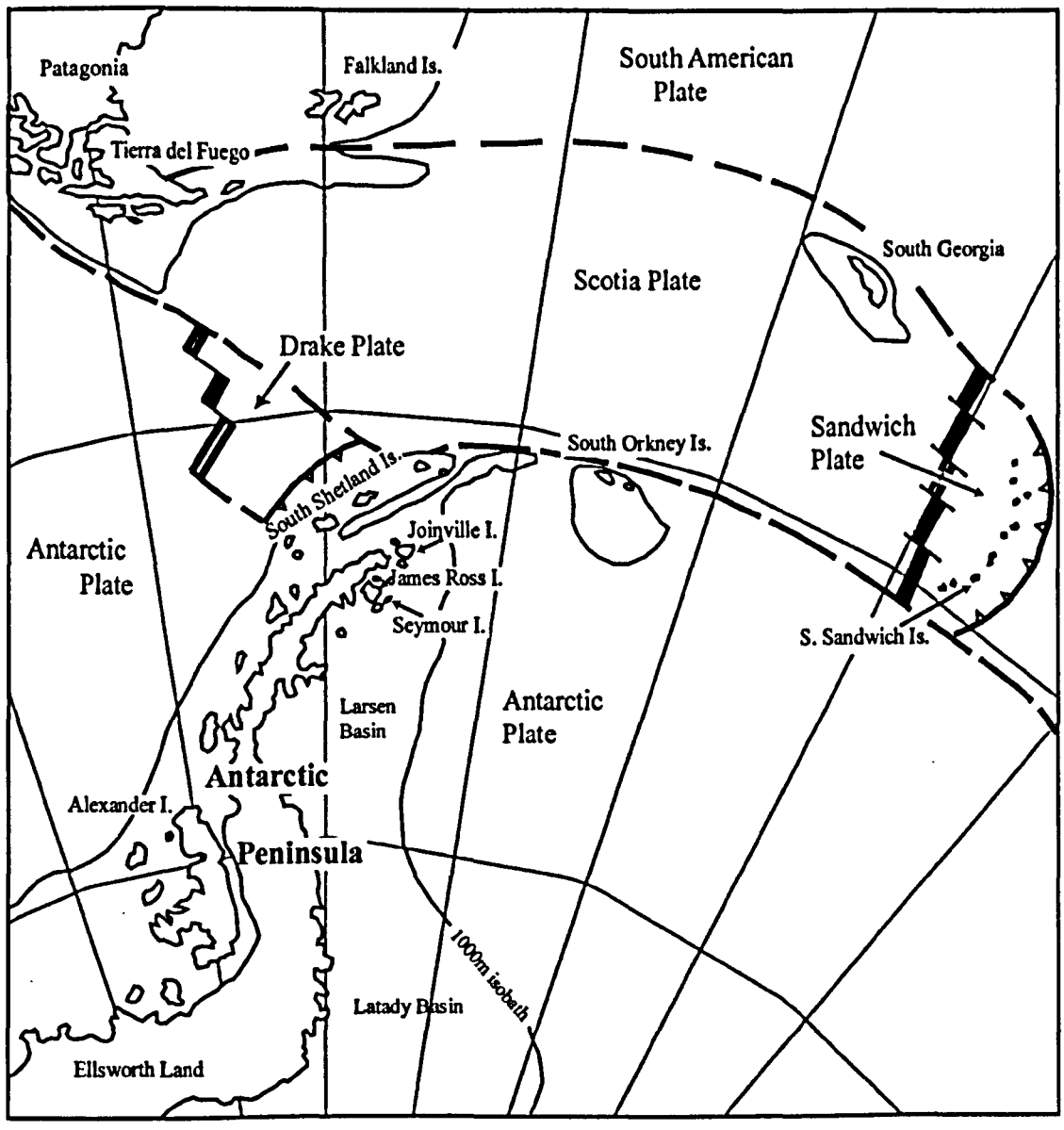


Fig. 1.2. Location Map of the Antarctic Peninsula region, with present geotectonic setting. Modified after Elliot (1988)

During the late Early and mid-Cretaceous, the deep marine sedimentation continued as further submarine fan conglomerates were deposited (Ineson, 1989). However, from the mid-Albian (Vaughan & Storey, 2000), the sediments begin to show evidence for uplift of the Antarctic Peninsula, as the facies represent deposition in shallower water conditions (Macdonald & Butterworth 1990). Tectonic uplift of the arc may be related to a period of terrane accretion (Vaughan & Storey, 2000) that badly deformed sediments in the southern part of the arc (Storey *et al.*, 1996). Sediments of the Larsen Basin show no such deformation.

Uplift rates declined by the end of the Cretaceous, and the Larsen Basin area underwent subsidence (Ineson, 1989). As a consequence, throughout the Late Cretaceous and the majority of the Paleocene there was essentially continuous deposition of sediments. Basin subsidence had declined in the James Ross Island area (Pirrie *et al.*, 1991) allowing sediments to prograde into the James Ross Basin and depositing shallow-marine facies (Hathway, 2000) throughout the Paleogene. Hathway (2000) also recognised the development that basin margins in the James Ross Island area had become fault-controlled from the Aptian onwards.

Oceanic cool water circulation began to develop around the southern edge of the Pacific Basin (Dingle & Lavelle, 2000) during the Maastrichtian to mid-Paleocene. The Drake Passage then opened between the southern tip of South America and the northern tip of the Antarctic Peninsula, while the Tasman Sea had opened between Antarctica and southern Australia, by the Late Eocene-Early Oligocene (Dingle & Lavelle, 2000). This allowed the development of deep water circulation around the entire continent for the first time. The thermal isolation of Antarctica led to high latitude cooling and the development of permanent ice on the continent. However, the sediments of the area do not record this transition due to a

region-wide unconformity (Cape Roberts Science Team, 2000). By the time that sedimentation resumed, as recorded by deposits in the fore-arc basins of the Antarctic Peninsula on King George Island, glacial conditions were already present.

1.3 - LITHOSTRATIGRAPHY OF THE JAMES ROSS BASIN

The Antarctic Peninsula region is the only part of Antarctica in which sediments of Late Cretaceous and Early Cenozoic age are exposed in near-continuous succession (Fig. 1.3). The greater part of the succession occurs on or around James Ross Island (Elliot 1988), which lies to the southeast of the Antarctic Peninsula in the James Ross Basin extension of the Larsen Basin.

The clastic sediments of the James Ross Basin are divided into three groups (Fig. 1.4). The basal sedimentary sequence of this basin, which is the Nordenskjöld Formation of Late Jurassic age (Farquharson 1982). The Early to Late Cretaceous Gustav Group is made up of the Lagrelius Point, Kotick Point, Whisky Bay and Hidden Lake formations. The Late Cretaceous to Early Paleocene Marambio Group is made up of the Santa Marta, the López de Bertodano and the Sobral formations, and the Paleogene Seymour Island Group made up of the Cross Valley and La Meseta formations. The sediments in the James Ross Basin are topped by the James Ross Island Volcanic Group of Miocene age (Pirrie et al. 1991).

1.3.1 - Nordenskjöld Formation

Lithology: The Nordenskjöld Formation sediments are comprised of alternating thin horizons of radiolarian-rich mudstones and tuffs (Elliot 1988), within which three members are recognised (Longing, Ameghino and Larsen).

Thickness: The thickness of the Nordenskjöld Formation is uncertain.

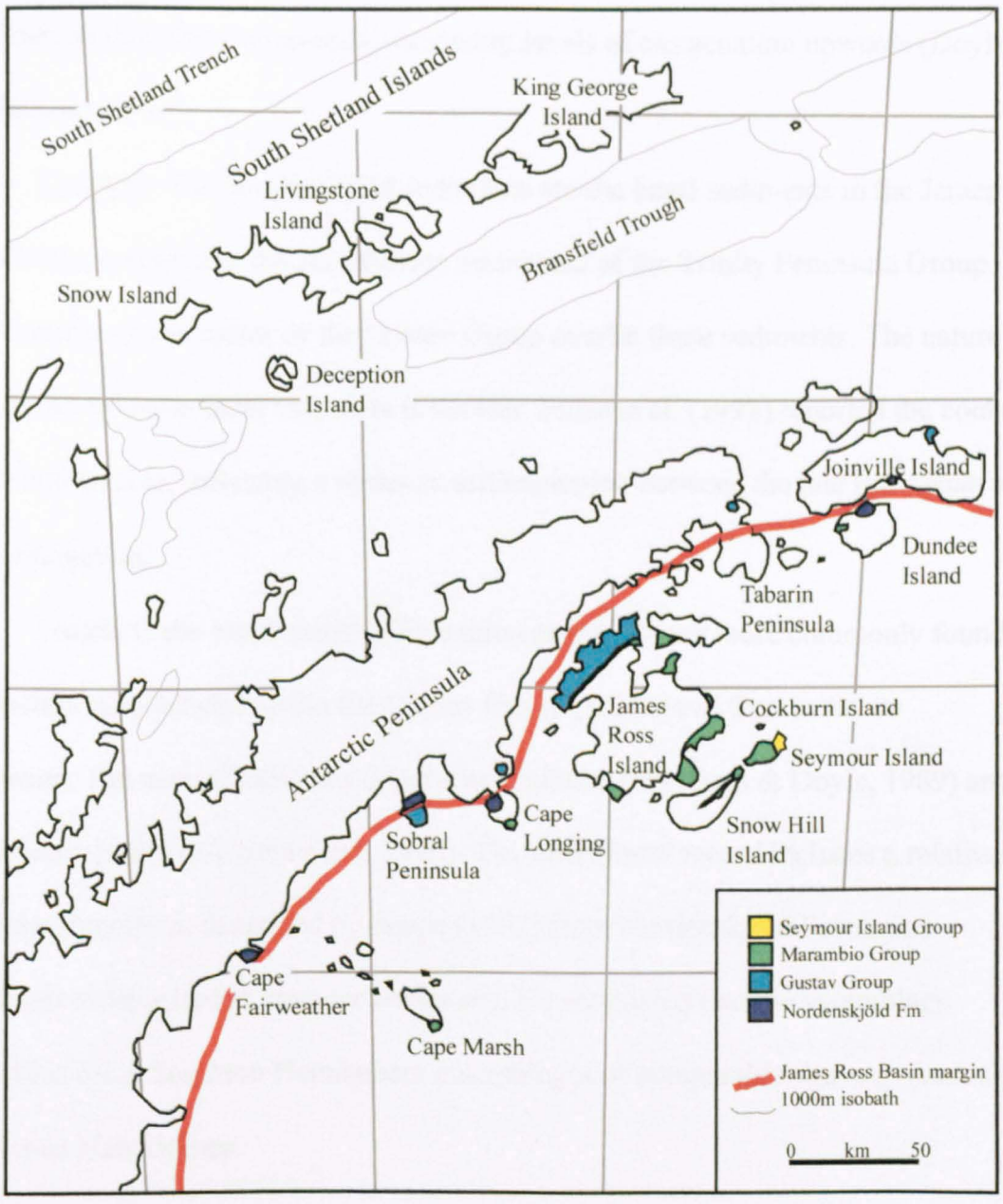


Fig. 1.3. Location and simplified geological map of the northern Antarctic Peninsula. Modified after Elliot (1988).

Depositional Environment: The sediments were deposited in the anoxic environment of a back-arc basin related to a volcanic arc (Farquharson 1982) positioned at the same site as today's peninsula. Transitions between the three members of the Nordenskjöld Formation are thought to be linked to an uplift of the sea floor, as they show apparently increasing levels of oxygenation upwards (Doyle & Whitham, 1991).

Contacts: The Nordenskjöld sediments are the basal sediments in the James Ross Basin, and overlie the accretionary sediments of the Trinity Peninsula Group. The Cretaceous sediments of the Gustav Group overlie these sediments. The nature of the contact between these two units is unclear. Pirrie et al. (1992) reported the contact as unconformable, reflecting a hiatus in sedimentation between the late Berriasian to the Valanginian.

Fossils in the Nordenskjöld Formation sediments are more commonly found in allochthonous blocks within the Gustav Group (see below) than in *in situ* sediments. The macrofossil record includes molluscs (Whitham & Doyle, 1989) and crustaceans (Aguirre-Urreta *et al.*, 1990). The microfossil record includes a relatively diverse palynoflora, described by Snape (1992) from Nordenskjöld Formation sediments redeposited in younger sediments. He concluded that the assemblage resembled other Southern Hemisphere assemblages of comparable age, e.g. Australia and Papua New Guinea.

Age: The sediments are dated to the Late Jurassic, Kimmeridgian - Tithonian, according to Elliot (1988).

Comments: The Nordenskjöld Formation outcrops in small, separate areas along the eastern edge of the peninsula at Cape Fairweather, the Sobral Peninsula, Cape Longing, Dundee Island and Joinville Island (Fig. 1.3). These *in situ* localities

have been subjected to localised thermal heating, and consequently are barren of palynomorphs, making them useless for palynological biostratigraphic analysis. However, they do contain invertebrate faunas that can be used for biostratigraphy. Other Nordenskjöld sediments are present as small clasts and large blocks (200 x 800m) (Pirrie *et al.* 1992) in younger Cretaceous sediments of the Gustav Group, and these sediments contain macrofossils and also microfossils that are of stratigraphic use.

1.3.2 - Gustav Group

The Gustav Group sediments occur predominantly on the northwest coast of James Ross Island with smaller outcrops on the Antarctic Peninsula. They are, for the most part, a conglomerate and sandstone sequence, although all of them contain volcanic debris and ash fall material. The Gustav Group represents the proximal deposition of submarine fan complexes and deep marine slope sediments that were derived from the volcanic arc now represented by the Antarctic Peninsula. The pattern of sediment deposition was controlled by an active fault on the western basin margin. This active fault margin, and related slope instability, resulted in large slide blocks of Nordenskjöld Formation being deposited into Gustav Group sediments. Towards the top of the Gustav Group, there is evidence of uplift from the basal continental slope to an outer shelf setting depth (Crame *et al.*, 1991).

- Lagrelius Point Formation

Lithology: The lowermost sediments are a pebble and cobble conglomerate sequence with some interbedded sandstones

Thickness: At least 500m thick

Depositional Environment: Submarine fan deposits

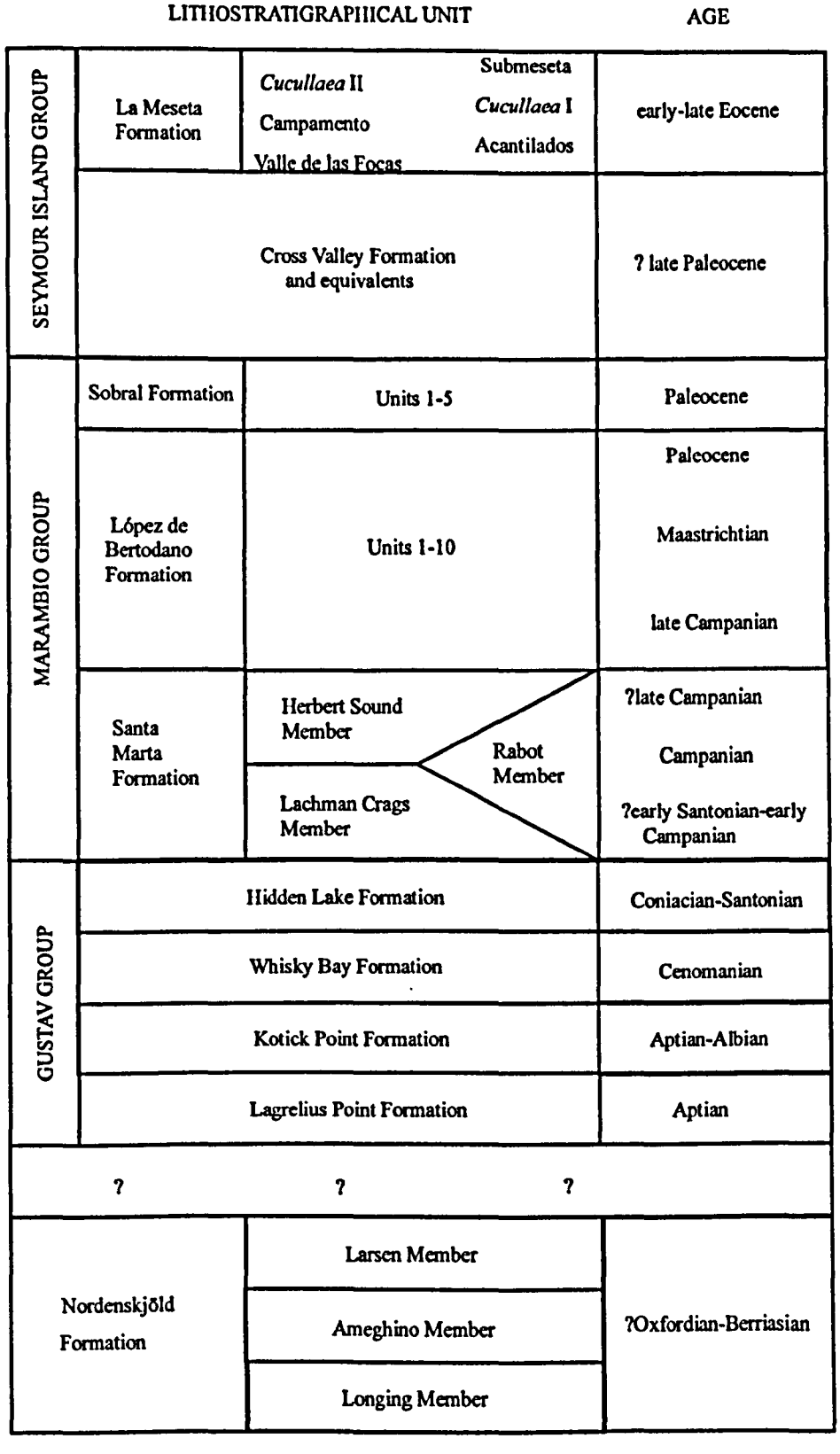


Fig. 1.4. Summary of the lithostratigraphy of James Ross Basin after Macellari (1988), Pirrie *et al.* (1992) and Marenssi *et al.* (1998)

Contacts: Overlies (probably unconformably) the Nordenskjöld Formation, and is overlain by the Kotick Point Formation; the nature of the contact is unclear.

Fossils: Palynomorphs (Ineson *et al.* 1986)

Age: Early Aptian (Riding *et al.*, 1998)

Comments: According to Keating *et al.* (1992, p.279) 'the formation forms steep, poorly accessible cliffs and its stratigraphical relationships are poorly known.'

- Kotick Point Formation

Lithology: The Kotick Point Formation is made up of horizons of interbedded breccia and conglomerates containing large clasts, many from the Nordenskjöld Formation., and thick intervals of medium-fine-grained sandstones, silty mudstones and clays.

Thickness: Uncertain.

Depositional Environment: Base of the continental slope, with both submarine fans and deep marine slope sediments.

Contacts: Overlies the Lagrelius Point Formation, and the nature of the contact between the two formations is unclear. The Kotick Point is in turn overlain by the Whisky Bay Formation, and again the nature of the contact is unclear.

Fossils: Ammonites (*Silesites*), belemnites and bivalves. Diverse palynofloras, both marine and terrestrial

Age Aptian-Albian (Ineson *et al.*, 1986).

- Whisky Bay Formation

Lithology: The Whisky Bay Formation shows a large amount of lateral variation, from mudstones to boulder conglomerates. It has been divided into six members by Ineson *et al.* (1986) in local areas, with the possibility of lateral continuation between these members.

Thickness: 2070-2370m.

Depositional Environment: Base of the continental slope, with both submarine fans and deep marine slope sediments.

Contact: Overlies the Kotick Point Formation, and is overlain by the Hidden Lake Formation. The nature of the contact is unclear.

Fossils: Bivalves, ammonites and palynofloras.

Age: Cenomanian (Riding *et al.*, 1998)

- Hidden Lake Formation

Lithology: The Hidden Lake Formation is comprised of a sequence of volcaniclastic sandstones interbedded with conglomerate horizons in the lower part of the formation, and siltstone horizons in the upper part of the formation (Elliot 1988).

Thickness: 400m.

Depositional Environment: The coarser-grained conglomeratic sediments in the lower part of the formation are held to represent deposition at the base of the continental slope (Crame *et al.*, 1991). The finer-grained sediments at the upper part of the formation represent a shallower environment of deposition in an outer shelf setting, below the storm wave base.

Contacts: Overlies the Whisky Bay Formation, and as the uppermost formation of the Gustav Group, is overlain by the Santa Marta Formation, the lowermost formation of the Marambio Group.

Fossils: A varied palynoflora, both terrestrial and marine, inoceramid bivalves.

Age: Coniacian-Santonian (Riding *et al.*, 1998)

Comments: The dinoflagellate flora indicates 'a nearshore coastal marine environment" (Barreda *et al.*, 1999, p. 53).

1.3.3 - Marambio Group

The sediments of the Marambio Group, the Santa Marta, López de Bertodano and Sobral formations, are more widely distributed than the underlying Gustav Group. They are found on the southeast part of James Ross Island, and on Snow Hill, Seymour, Cockburn and Cape Marsh Islands (Fig. 1.3). These sediments are primarily fine-grained clastics, deposited in the proximal part of the basin as the rate of uplift of the arc declined (Elliot 1988). The sediments of the Marambio Group have been affected by sea level changes through the Danian, moving from a shelf environment below the wave base to a wave-dominated setting

- Santa Marta Formation

Lithology: The Santa Marta Formation is made up of silty and muddy sandstones with interbedded conglomerates, marls, calcite-cemented beds and concretionary beds.

Thickness: ?1000m.

Depositional Environment: Mid outer shelf and inner shelf environments (Pirrie, 1989, 1990).

Contacts: Nature of the contacts in unclear.

Fossils: Ammonites, bivalves, palynofloras.

Age: Santonian - Campanian.

- López de Bertodano Formation

Lithology: The López de Bertodano consists of fine sands and sandy siltstones with occasional carbonate-cemented sandstones. Concretionary beds are again seen, occasionally abundant in parts of the sequence. Upper parts of the section contain volcaniclastic sediment and glauconite beds.

Thickness: 1190m.

Depositional Environment: Shallow marine shelf, mostly below wave base.

Contacts: Nature of the contacts in unclear.

Fossils: ammonites, palynomorphs, bivalves, gastropods, echinoids, corals, annelids, crustaceans, bony fish, sharks, mosasaurs and plesiosaurs.

Age: Campanian - Paleocene.

Comments: The Cretaceous-Tertiary boundary is found in the upper part of the Formation (Askin, 1988b), and the uppermost sediments extend into the Danian. During the Danian, the area underwent uplift, or regression, and the Seymour Island area was brought closer to or above sea level (see p. 7).

- Sobral Formation

Lithology: The sediments are predominantly silts, and sandstones. The Formation has a basal silty section that becomes coarser and dominated by glauconite sandstones up-section.

Thickness: 255m.

Depositional Environment: Initially shallow marine, later sediments deposited on a shallow shelf in a wave-dominated environment (Elliot, 1988).

Contacts: Lies unconformably over the López de Bertodano Formation.

Fossils: Palynomorphs, bivalves, bony fish, sharks.

Age: Paleocene.

Comments: A transgression, or basin subsidence, allowed sedimentation to begin again after the unconformity at the top of the López de Bertodano Formation, although the abundant glauconite suggests slow deposition (Elliot, 1988). Towards the top of the section, a further regression brought the sediments back to the level of the wave base.

1.3.4 - Seymour Island Group

The Seymour Island Group is almost completely restricted to Seymour Island sections, although basal La Meseta sediments also crop out on Cockburn Island. The area was subject to tectonic activity, related to the evolution of the Antarctic Peninsula, during the Late Palaeocene. A fault-bounded valley developed through the existing Marambio Group sediments into which the lower Seymour Island Group sediments were deposited. This pattern was repeated again in the Eocene.

The Seymour Island Groups sediments continue this shallowing trend, and are thought to have been deposited at, or near to, the shoreline.

- Cross Valley Formation

Lithology: The basal Seymour Island unit, the Cross Valley Formation, is a small unit of coarse sands and conglomerates of little differentiation, which have a distinct volcanic element over which lie horizons of sandstone.

Thickness: 105m.

Depositional Environment: The sediments were deposited in a canyon, presumably carved during a marine regression, or uplift. The sediments are thought to have been deposited rapidly as there is little sorting or evidence of bedding (Elliot, 1988). The setting may be marine due to the presence of marine invertebrates.

Contacts: Lies unconformably on the Sobral Formation.

Fossils: Marine invertebrates. Palynomorphs recovered from clasts.

Age: (?)Late Paleocene.

Comments: The Cross Valley Formation is a small, localised unit in the north east of Seymour Island. Porebski (2000) links its formation to fault subsidence, and describes it as accumulating in an incised shelf valley. Tectonic activity was

widespread at this time, with large amounts of volcaniclastic sediment present, as well as evidence of uplift and tilting. The younger sediments of the Cross Valley appear to be contemporaneous with sediments of the Sobral Formation, suggesting that while part of the basin had been uplifted and incised, deposition continued as before throughout the rest of the area, albeit with an increase in the amount of volcanics being deposited (D. Cantrill, pers. comm.).

- La Meseta Formation

Lithology: Unconsolidated silts and sands with occasional cemented sandstone horizons that show 'large-scale coarsening-upward cycles from mudstones to sandstones, with laterally complex shell beds at the top of the cycles' (Pirrie et al. 1992 p.263).

Thickness: 750m.

Depositional Environment: Shallow marine. Evidence for transgressive-regressive cycles occur throughout the formation (Marenssi *et al.*, 1998). The sediments have been deposited within a fault-bounded incised valley estuary (Porebski, 2000).

Contacts: Lies unconformably over the López de Bertodano and Sobral Formations.

Fossils: Highly fossiliferous. Contains palynomorphs, bivalves (which form large coquinas), bony fish, sharks' teeth, penguins and other vertebrates (Sadler, 1988).

Age: Eocene.

Comments: The La Meseta Formation represents a marine regresssion, with sediments, including a large terrestrial component, being deposited near sea level.

There is no evidence for volcaniclastic sedimentation in the sequence, indicating a decline in subduction and related volcanism.

1.4 - SEYMOUR ISLAND

The field area for this thesis is on Seymour Island. Seymour Island lies to the southeast of James Ross Island and the northern tip of the Antarctic Peninsula (Fig. 1.3) at a latitude of 64°15'S. and longitude 56°45'W. It comprises sediments from both the Marambio Group (López de Bertodano and Sobral formations) and the Seymour Island Group (Cross Valley and La Meseta formations). These sediments range from Maastrichtian to Late Eocene in age (see above for description of general stratigraphy and deposition). The sequences contain many distinctive lithologic marker bands in the Sobral and La Meseta formations, and are also highly fossiliferous, which allows good biostratigraphic correlation and palaeoenvironmental studies.

The formations outcropping on Seymour Island are described in further detail below:

- López de Bertodano Formation

The López de Bertodano Formation was first defined by Rinaldi *et al.* (1978). It comprises unconsolidated muddy sandy silts, and this remains quite consistent through the formation according to Macellari (1988). The López de Bertodano Formation however, does show an increase in the proportion of volcanic-derived material and glauconite towards the upper part of the section (Macellari, 1988). Subdivided into 10 smaller members based on lithology and fossil content, the formation ranges in age from the Late Campanian to the earliest Danian (Macellari, 1988).

Shallow-marine, fossiliferous, transgressive sequence. Silty muds, with increasing occurrence of volcanism and glauconite in the upper beds.

- Sobral Formation

The Sobral Formation is separated from the underlying López de Bertodano Formation by an erosional unconformity. The lithology is more varied than in the earlier sediments, and contains not only silty mudstones and glauconite (as in the sediments mentioned above), but also tuffaceous and other volcanic sediments, and sandstones of varying coarseness. The lower part of the formation is silty mudstone, dominated by volcanic-derived sediment and glauconite, and is similar to the upper part of the López de Bertodano Formation. Later sediments show an increase in the proportion of quartz-rich sandstones. Towards the top of the Sobral Formation, the sediments become increasingly coarse-grained, are cross-bedded, and again show an increase in the amount of glauconite and volcanic content present.

The Sobral Formation has previously been included in other stratigraphic groups, as well as being described in separate formations, as reviewed in Macellari (1988). It is currently subdivided into 5 members (Sadler, 1988), and is considered earliest Paleocene (Danian) in age.

- Cross Valley Formation

The Cross Valley Formation was first defined in 1982 by Elliot & Trautman, as a unit of sedimentary facies that cuts though the northern part of the Sobral Formation. The lithology of the formation, most notably in the lower part of the section, is dominated by poorly consolidated, coarse-grained sandstones, stained by limonite (Sadler, 1988). Higher in the section are further sandstones and siltstones, although with less marked staining, and then coarser sediments appear, with very coarse sandstones and conglomerates (Sadler, 1988). The conglomerates reflect a

variety of sources, such as volcanic tuffs, lacustrine freshwater mudstones and probable fluviatile sandstone clasts (D. Cantrill, pers. comm.). The uppermost part of the Cross Valley Formation is made up of facies seen lower down in the section, as well as a further distinct fossiliferous shale facies. This latter facies was reported by Zinsmeister (in Sadler, 1988) to be conformable with beds of the Sobral Formation.

The Cross Valley Formation is recognised (Sadler, 1988) as the sedimentary fill of a channel that was formed in the underlying Sobral Formation by the action of an underlying fault system (Porebski, 2000). According to Sadler (1988) this erosional event occurred while deposition of the Sobral Formation was still ongoing, as in parts of the outcrop it is observed to be conformable with the Sobral.

- La Meseta Formation

The La Meseta Formation, as with the Cross Valley Formation, was originally named and described by Elliot & Trautman (1982). It was named for the meseta that stands on the northern part of the island, and around which the formation crops out. The La Meseta Formation sediments were deposited into an estuary system, which had been initiated in underlying Late Cretaceous and Paleocene sediments of Seymour Island by local subsidence along pre-existing faults (Porebski, 2000).

The dominant lithologies of the La Meseta Formation are silty sands and muds, which are present as the infill of various channels which cross-cut each other. Each of these channel units has an erosional base, with coarse sandstones and conglomerates, which fine-upwards into finer-grained sediments and are topped with shell coquinas comprising mostly *Cucullacea* shells and shell fragments (Marenssi *et al.*, 1998).

1.5 - PALYNOLOGY

The term palynology was first coined by Hyde & Williams (1944) in order to distinguish the young science from 'pollen analysis', pioneered by von Post in Sweden (1917) and which deals with Quaternary spores and pollen. Palynology is the study of the organic constituents, or palynomorphs, that survive a vigorous acid maceration process to leave an acid-insoluble residue. These constituents include chitinozoans, acritarchs, foraminifera linings, fungal spores, plant fragments, algal cysts (including dinoflagellate cysts) and plant isospores, microspores, megaspores and pollen grains (isospores, microspores, small megaspores and pollen grains are collectively known as miospores).

The principle use of palynomorphs has been for their biostratigraphical information, such as being able to recognise the palynomorphs that characterise a particular stratigraphical position in the geological column. This information can then be used to ascribe a relative age for the strata and thus correlate it with other strata in different areas. Characterising the entire palynoflora, which includes all acid insoluble debris including amorphous organic matter, is also useful to determine the depositional environment.

Marine palynofloras are dominated by algal cysts. Some are of uncertain origin, and are included in the group 'Acritarcha'. The Acritarcha is an informal group that is used when algal cysts cannot be assigned to any known group. The majority of the palynoflora, from the Mesozoic onward, are dinoflagellate cysts. It is this group that is the focus of Mesozoic and Cenozoic marine palynology due to their rapid evolution, and their consequent use in biostratigraphical studies.

The terrestrial palynomorph flora is made up of spores (produced by bryophytes and vascular cryptogams or pteridophytes), gymnosperm and angiosperm pollen. All three groups are composed of sporopollenin, an organic compound that resists decay and so protects the contents of the spore/pollen. All three groups have a different morphology, and are produced by different life-cycles. This affects the distribution of the parent plant and thus also the produced spores and pollen.

Since the 1970s, e.g. Wall *et al.* (1977), research has been carried out into other aspects of palynomorph distribution, such as their use as palaeoenvironmental indicators (see Chapter 4). They are also useful in that having both a marine and a terrestrial component allows for the correlation between offshore and onshore sequences.

1.6 - ANTARCTIC PALYNOLOGY

The first terrestrial plant fossils to be collected in Antarctica were those recovered from Seymour Island by the Norwegian Captain C.A. Larsen during the 1892-1893 season (Zinsmeister, 1988). He had been sent south on a whaling expedition following the reports of Sir James Clark Ross regarding the quantities of whales in the Southern Ocean. While in Antarctica, he landed on Seymour Island and collected fossil shells and wood from the beaches, material that was identified on his return as being from the Lower Tertiary.

The publication of field notes of the visit to Seymour Island, together with details of the fossils discovered, encouraged the formation of the Swedish South Polar Expedition, led by Otto Nordenskjöld. Lasting from 1901-1903, the Expedition retrieved a variety of material, including ammonites, varied leaf impressions, penguin

bones and rock samples. Following the end of this visit, little effort was made to collect further material for many years.

Between 1908 and 1912, papers were published on the plant fossils collected by Nordenskjöld, but not until 1959 was anything published on the plant microfossils. In what was one of the first palynological papers published on Antarctic material, Cranwell (1959) described fossil pollen from Seymour Island using material collected by the earlier Swedish Expedition. *Nothofagus* and conifers (mainly podocarpaceous and araucarian pollen) dominated the flora she described, with occasional pteridophyte spores. Cranwell recognised that the combination of floral communities was similar to other southern hemisphere localities, and represented different communities at different elevations. The age of the sediments were described as Early to Mid-Campanian in the south-west of the island, with younger sediments of Early Miocene age in the north-east. There were not thought to be any sediments of Paleogene age.

Cranwell *et al.* (1960), noted that outside of the northern Antarctic Peninsula no Cretaceous or Tertiary sediments had been discovered throughout the rest of Antarctica, and they carried out an investigation into glacial erratics found among the moraines of McMurdo Sound. The erratics in question were chosen because of their resemblance to distinctive sediments deposited in New Zealand at a time when the two areas were in close proximity. The sediments produced palynofloras with both a terrestrial and marine component, that were dated to Eocene or Oligocene in age, based on the foraminifera present. Dinoflagellates and chorate dinoflagellate cysts that Cranwell *et al.* (1960) termed 'hystrichosphaerids' dominated the marine flora with smaller numbers of microforaminifera and scolecodonts present. They interpreted the environment as 'being offshore of normal salinity and low turbidity'

(Cranwell *et al.* 1960 p.701). The terrestrial flora was not as abundant, containing taxa similar to those which had been previously described from Seymour Island (Cranwell, 1959). It also contained forms suggesting a strong South American character, although the flora still had recognisable links to other parts of the former Gondwana landmass. The terrestrial taxa were dated to being between the Late Cretaceous and the Oligocene age, it being impossible to refine the dating any further at that time.

A further analysis of the McMurdo erratics (Cranwell, 1964) allowed the hystrichosphaerids of Cranwell *et al.* (1960) to be identified as *Cordospheridium (Hystrichosphaeridium) diktyplokus* (Klumpp, 1953) Eisenack 1963. Cranwell (1964) recognised the significance of this species as a biostratigraphic marker, given its apparent narrow time range (Early Tertiary) and its distribution in other, similar sections in the Southern Hemisphere.

Subsequently, further work was carried out on Seymour Island due to the recognition that it contains the only outcrop with sediments of Late Cretaceous and Early Cenozoic age in Antarctica. Hall (1977) refined the dating of the different formations on Seymour Island; Late Cretaceous - Paleocene for the Cape Wiman unit (thought to be contemporaneous with the Sobral Formation); the Sobral Formation was placed in the Paleocene, and the younger La Meseta Formation was considered Late Eocene - Early Oligocene. No definite age could be deduced for the Cross Valley Formation.

Palamarczuk (1982) suggested a Danian age for the Sobral Formation from a single sample, noting the stressed-marine, and probably estuarine conditions that it represented, being dominated by *Palaeoperidinium pyrophorum* (Ehrenberg 1838) Sarjeant 1967b. Palamarczuk *et al.* (1984) later studied both the López de Bertodano

and Sobral formations, identifying the López de Bertodano Formation as being Campanian in age and the Sobral Formation as Danian, the two formations being separated by an erosional unconformity. This led to the erroneous conclusion that the Maastrichtian was not represented on Seymour Island, and that the Cretaceous-Tertiary Boundary (K-T) interval was not present. This assumption was subsequently disproved by Askin (1988b), who identified Maastrichtian age sediments and the K-T Boundary section in the upper part of the López de Bertodano Formation.

The first palynomorph zonation scheme for Seymour Island was compiled by Askin (1988a). It contained eight zones, and included samples from the James Ross Basin, not just Seymour Island. This was supplemented by a more detailed scheme covering the youngest Seymour Island sediments, the Cross Valley and La Meseta formations, by Wrenn & Hart (1988).

The La Meseta Formation contains some of the youngest outcropping sediments in the Larsen Basin (except for the Miocene Hobbs Glacier Formation, the James Ross Island Volcanics Group and the capping Holocene glacial Weddell Formation sediments that overlie the La Meseta Formation). There is a significant unconformity between the upper beds of the La Meseta Formation and the following Oligocene beds of King George Island. King George Island is part of the Southern Shetland Islands, lying to the west of the Antarctic Peninsula, in what was an intra-arc basin setting. The sediments of King George Island contain plant macro- and microfossils, predominantly *Nothofagus* and conifer (podocarpaceous and araucarian). The sediments themselves are volcanogenic sandstones and ashes deposited in a variety of terrestrial (intra-arc) settings (D. Cantrill, pers. comm).

By the 1980's, the importance of the Cenozoic record in the high latitudes became clear, as understanding grew of how the geographical isolation of the

continent has dominated global climate change evolution during the Cenozoic. Efforts were made by co-operating international teams to drill cores to recover sediments that would allow assessment of the timing of the onset of glaciation, to bridge the gap between Eocene and Oligocene sediments, and so determine the pattern of climate deterioration on the continent.

Three drilling projects were set up during the 1980's and 1990's. The first being MSSTS-1 (1985), the second CIROS-1 (1988), and the third and most recent being the Cape Roberts Drilling Project (1997-1999). All three drill sites were in the Ross Sea, on the west coast of the McMurdo Sound (part of East Antarctica), near the Antarctic bases of McMurdo and Scott (Cape Roberts Science Team 1998). The sediments drilled were a seaward -dipping sequence of over 2000m, and the aim was to eventually recover a continuous core, in overlapping sections, through the strata from 30 Ma to 100 Ma (Oligocene to Albian) (Cape Roberts Science Team, 1998).

The palynological flora recovered from the MSSTS-1 site was Oligocene - Miocene in age, The *in-situ* flora was relatively sparse, but the sediments also contained a reworked Eocene flora, containing many distinctive species (Truswell, 1986). The CIROS-1 borehole recovered sediments that were dated to the Early Oligocene by both marine and terrestrial floras (Mildenhall, 1989; Wilson, 1989). These sediments also contained numbers of reworked Eocene and Paleocene species.

The Cape Roberts Drilling Project failed in its attempt to reach Paleogene sediments before reaching the basement. The upper part of the third, and deepest core to be drilled (CRP-3), contained reworked or unknown taxa, predominantly peridinioid, that could not be used as biostratigraphic indicators. Other microfossils groups (nannofossils and diatoms) dated this part of the section as Early Oligocene. No older sediments than this could be dated due to a paucity of time-diagnostic

fossils, however, the unknown taxa recorded from the upper part of the section were present throughout the entire section. This suggests that the sediments are of similar age throughout the core (Cape Roberts Science Team, 2000).

Studies of Holocene sediments in the southern Indian Ocean (Marret & de Vernal, 1997) and in the southern Atlantic Ocean and Weddell Sea (Harland *et al.* 1998) have also produced distinctive dinoflagellate floras. This, in combination with the studies described above, provides evidence of dinoflagellate cyst floras in Antarctica throughout the Cenozoic.

Chapter 2 - MATERIALS AND METHODS

2.1 - LOCALITIES

The location of all sections and spot samples in the López de Bertodano, Sobral, Cross Valley and La Meseta formations used in this thesis are marked on Fig. 2.1. which shows their topographic positions.

2.1.1 - López de Bertodano Formation

The sediments of the López de Bertodano Formation, which outcrops on the southwestern part of Seymour Island (see Chapter 1), are described as being 'monotonous' muddy sandy silts (Macellari, 1988, p. 26) with few sedimentary structures to allow a thorough examination of depositional environments. Consequently, the López de Bertodano Formation has been divided into 10 units based on faunal, physiographic and lithologic changes (Macellari, 1988) which can be difficult to discern when in the field. The Cretaceous-Tertiary Boundary (K-T) (Fig. 2.2) lies between units 9 and 10. Unit 10, the only part of the formation to be processed here is relatively sand-rich. The sand content increases further up-section, while concretionary horizons and glauconite also become common.

The samples collected (DJ 904) were part of a larger section (see Fig. 2.2) that covers Maastrichtian-Danian sediments across the Cretaceous-Tertiary Boundary. The samples were collected by Dr D. Cantrill (BAS) and others in the field season of 1999/2000. Only samples above the K-T Boundary were processed as part of this thesis, of which there were 34 from the López de Bertodano Formation (Fig. 2.3).

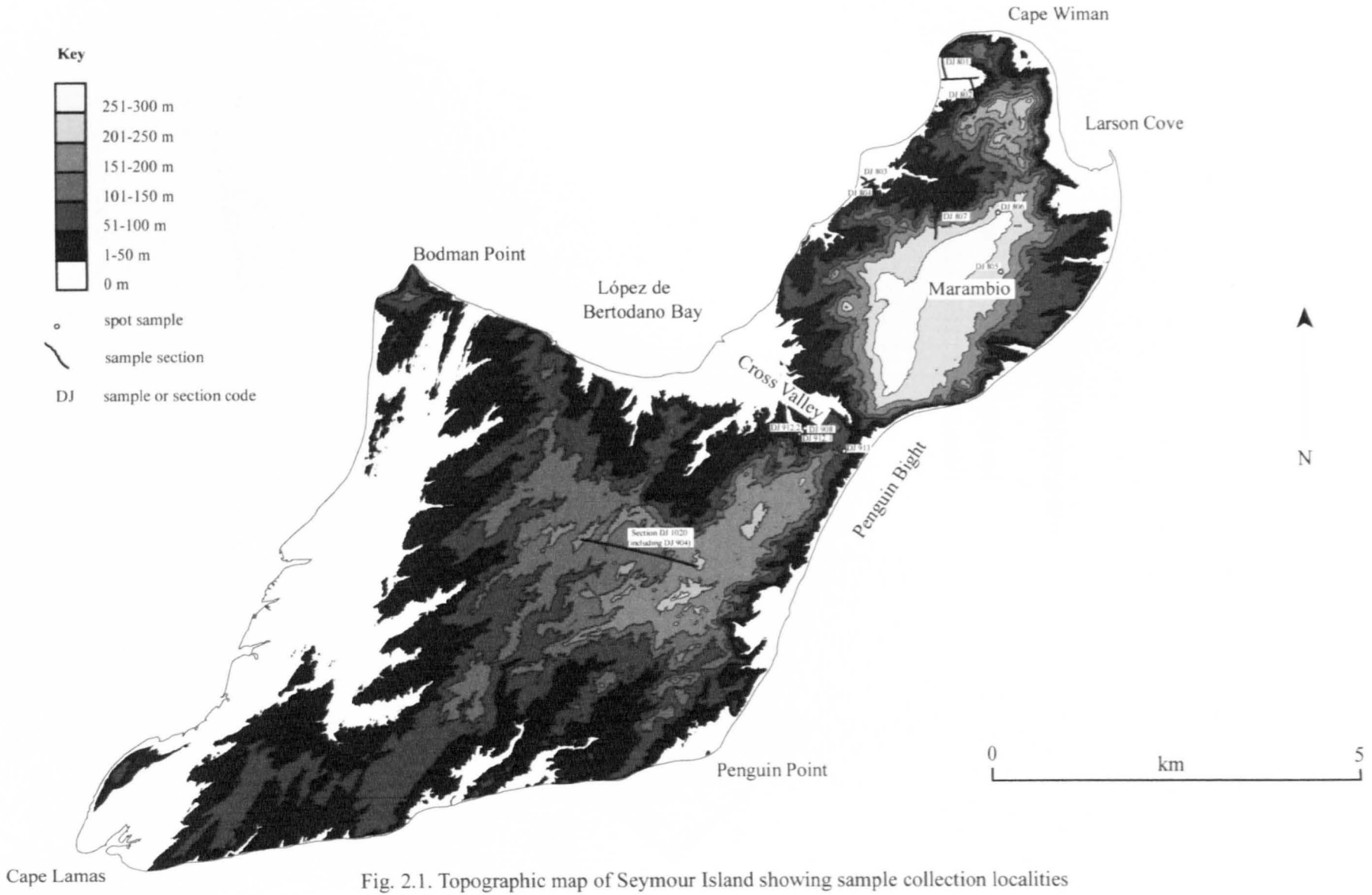


Fig. 2.1. Topographic map of Seymour Island showing sample collection localities

Fig. 2.2 Geologic map of southwest Seymour Island, showing location of Paleocene outcrop localities. After Harwood (1988, Fig. 1).

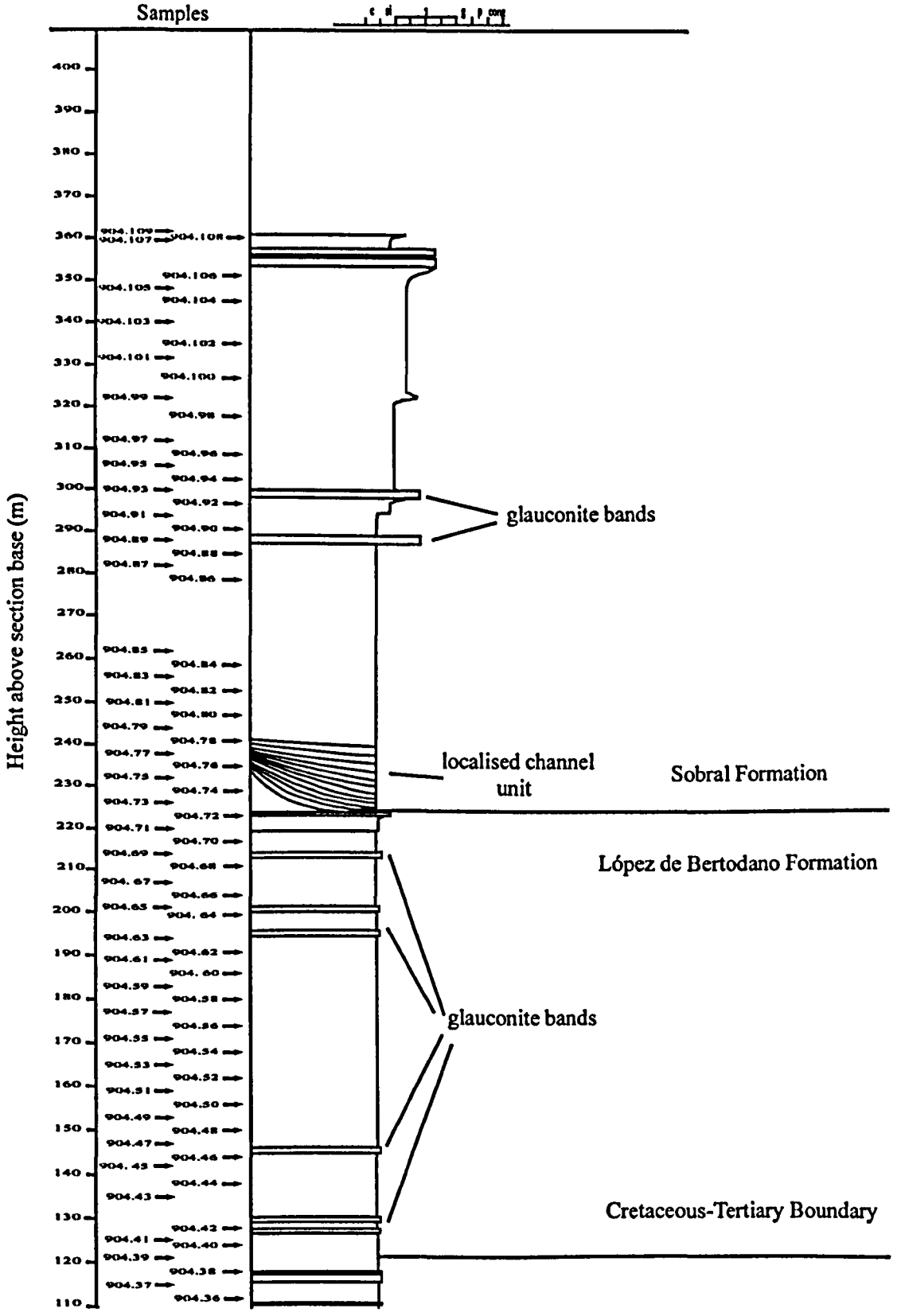


Fig. 2.3 Stratigraphic log of López de Bertodano and Sobral formations, with sample positions. Source: D. Cantrill, unpublished field data.

2.1.2 - Sobral Formation

The unconformity that separates the Sobral Formation from the underlying López de Bertodano Formation is erosive and downcuts up to 100m, in places it is channelised (D. Cantrill, pers. comm.). The amount of relief varies across the outcrop. The sediments of the Sobral Formation are have been divided into 5 units (Sadler, 1988) and show lenticular stratified sediments which are predominantly fine sandstones and silty mudstones (Sadler, 1988) which coarsen upwards to more glauconite-rich sandstones.

The Sobral Formation samples were collected as part of DJ 904 (see above). Thirty-seven samples were collected and processed from the base of the Sobral Formation (Fig. 2.3) where it unconformably overlies the López de Bertodano Formation.

2.1.3 - Cross Valley Formation

The Cross Valley Formation is a unit of sediment deposited within a channel carved into the Sobral Formation (Fig. 2.2). The sediments are predominantly coarse sandstones, although there are also conglomerate and siltstone horizons.

The Cross Valley samples were collected as spot samples (Fig. 2.2), not on a section during the 1999/2000 season. Sample DJ 908 is a clast of material from the base of the Cross Valley Formation (D. Cantrill, pers.comm.). Samples DJ 912.1-2 and DJ 913 were collected from stratigraphically higher in the formation.

2.1.4 - La Meseta Formation

The Eocene La Meseta Formation outcrops only on Seymour Island (1.3.1), and is the only Eocene sequence in the James Ross Basin. The Formation is made up

of transgressive-regressive, marine-estuarine cycle units (Porebski, 2000), which vary greatly in thickness and lateral extent. Consequently, no single vertical section can incorporate all the different units, and so it was necessary that a number of smaller, overlapping sections be sampled to ensure complete coverage (Fig. 2.4). Drs Dingle and Lavelle collected the 85 samples used in this work (DJ-801 - 807, Fig. 2.5) from the field during the field season of January - March 1996. The samples were collected as part of the High Latitude Palaeoclimates Project (HILATS), within the wider scope of the BAS Antarctic Palaeoenvironmental Change Programme (APECS).

The La Meseta Formation has been stratigraphically described and classified four times: initially by Elliot & Trautman (1977), then by Sadler (1988), who broke it down into a sequence of seven TELM units which were thought to occur within a delta. In 1995, both Porebski and Marenssi redescribed the La Meseta Formation. Porebski by simplifying it into three major units of fill, and Marenssi, who described six units bounded by unconformities and shell coquinas, and who described the setting of the La Meseta Formation as being that of an incised valley estuary. Further analysis by Porebski (2000) refined this model by describing the valley as having been bounded by growth faults, thus explaining the La Meseta as a compound fill produced by repeated downfaulting of the valley floor, increasing the accommodation space of the valley.

The classification of Marenssi (1995) is used herein. The members of the La Meseta Formation are the basal Valle de las Focas, the Acantilados, the Campamento, *Cucullaea* I, *Cucullaea* II and the uppermost Submeseta.

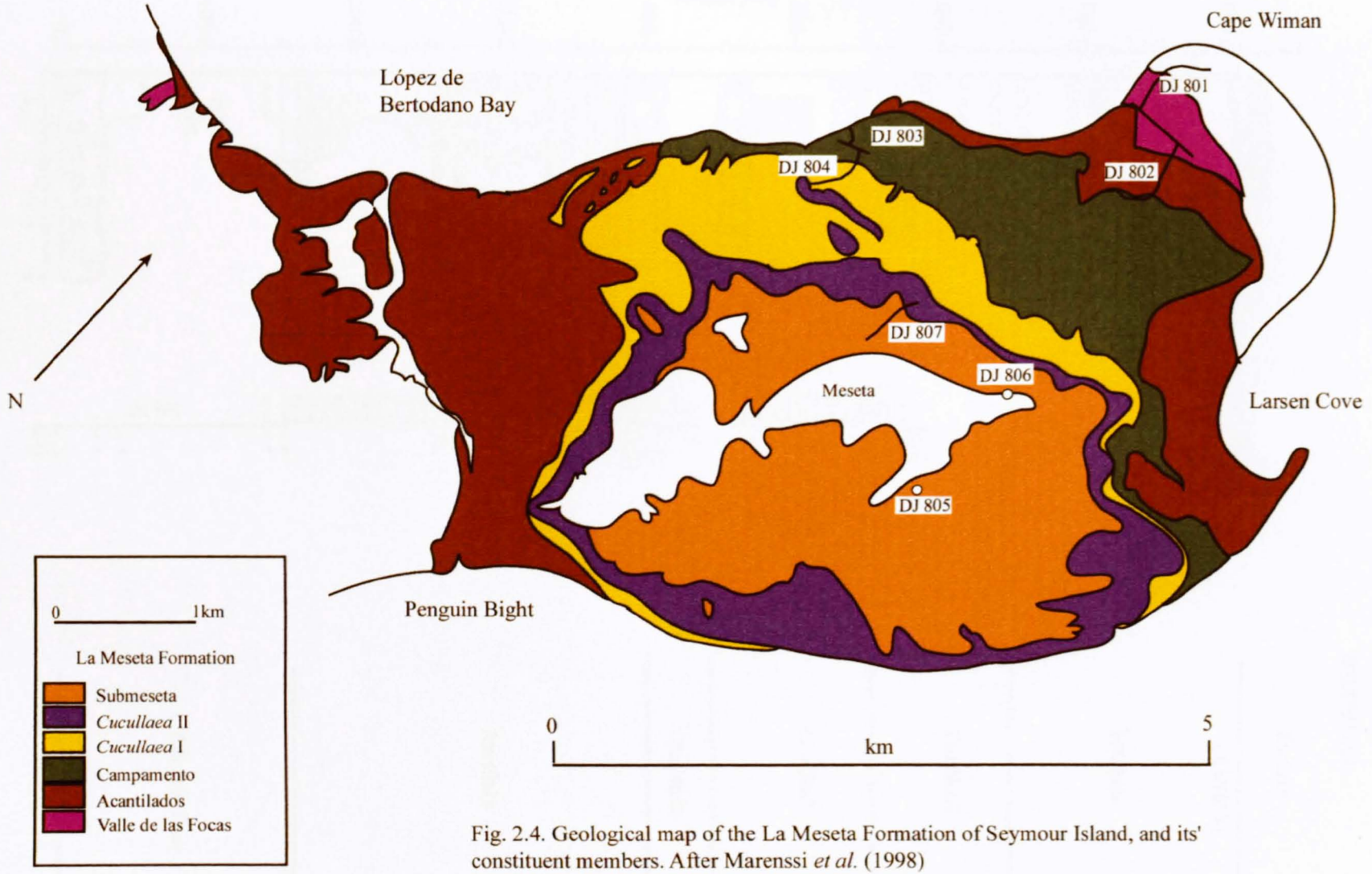


Fig. 2.4. Geological map of the La Meseta Formation of Seymour Island, and its' constituent members. After Marenssi *et al.* (1998)

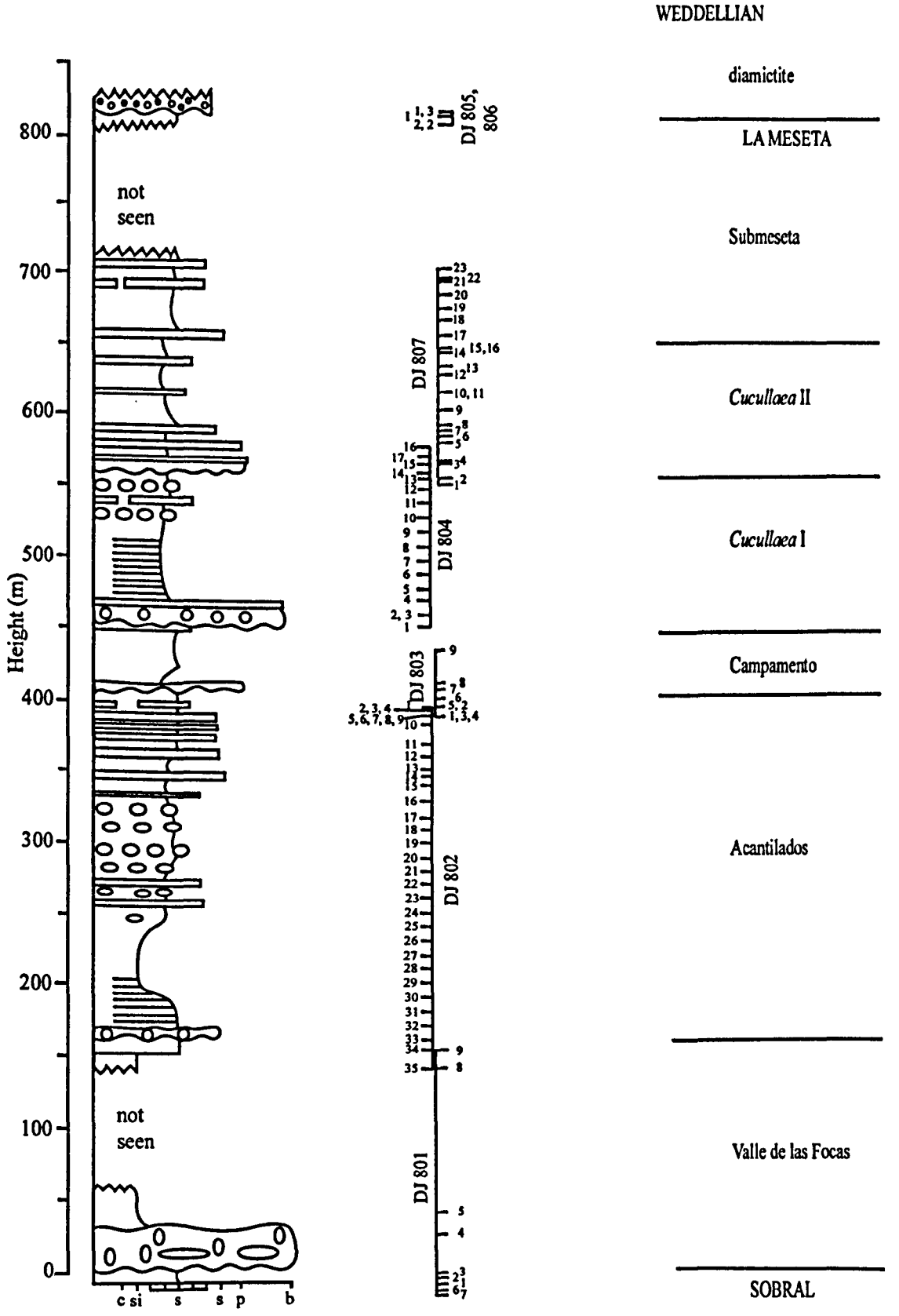


Fig. 2.5. Stratigraphic log of the La Meseta Formation with sample positions

2.2 - PROCESSING

The majority of palynological studies rely on counting a certain number of palynomorphs (usually 300) to provide a statistically relevant data-set. Once this has been achieved then a biostratigraphic study can be carried out as well as a palaeoenvironmental analysis based on the data. The use of absolute abundance counts i.e. the absolute abundance of dinoflagellate cysts or terrestrial spores and pollen per unit mass or volume of sediment gives an opportunity for the development of a more quantitative, as well as a more detailed data set. This is potentially of great value when trying to determine high-resolution events, as it allows the generation of a data-set that takes into account changes in abundance of the entire flora, rather than relative change within the flora. Absolute abundance counts are rarely carried out on pre-Quaternary sediments due to difficulty of determining accurate counts (Stover *et al.* 1996).

Different methods have been developed to ascertain absolute abundances of palynomorphs within a sample. The most common is the spiking of the sample by the addition of exotic markers before processing e.g. *Lycopodium* spore tablets, as it allows the calculation of pollen density (Stockmarr, 1971). Another method is to process a known amount of sediment, withdraw a known amount of residue at the end of the processing and then to count everything within that residue (Harland, 1989).

This study has been carried out using the method of Harland (1989) as it was felt that the data produced would provide an insight into the high-resolution palaeoenvironmental changes occurring in the field area (J. Powell, pers. comm.). This would be by not only analysing species changes through the section, but also by looking at the absolute numbers of the population. The drawback in using this method

is to ensure that as little material is lost as possible once the sample has been weighed. Consequently great care was taken over the processing at all times e.g. during the acid maceration process, excess liquid was removed using a pipette, rather than the decanting more frequently used, to ensure that the sediments were not disturbed and material lost. The samples were not oxidised as this can systematically affect the palynomorph population by the preferential removal of certain cyst species (Hopkins & McCarthy, 2000)

Processing was carried out as in Harland (1989), in order to calculate the number of cysts per gram of sediment. The methodology is given in detail below, while the basic processing procedure is summarised in Fig. 2.6: -

- 5 grams of sediment were dried in an oven overnight at 40°C
- The dried sample was reweighed, and the dry weight recorded as *a*
- Acid digestion in concentrated hydrochloric acid (to remove carbonates), followed by concentrated hydrofluoric acid (to remove silicates). Adding distilled water, which was then pipetted off after the sediment had settled, diluted the acids. This was repeated four times to lower the pH sufficiently to safely move to the next stage
- Further concentrated hydrochloric acid was added to remove insoluble ions and the sediment was then washed with distilled water
- The sample was then put through a heavy liquid separation using sodium polytungstate (specific gravity 2.2). This concentrated the organic fraction
- The organic residue was then split, half was left for palynofacies analysis, while the second half was stained with safranin, then sieved through 10μm mesh.

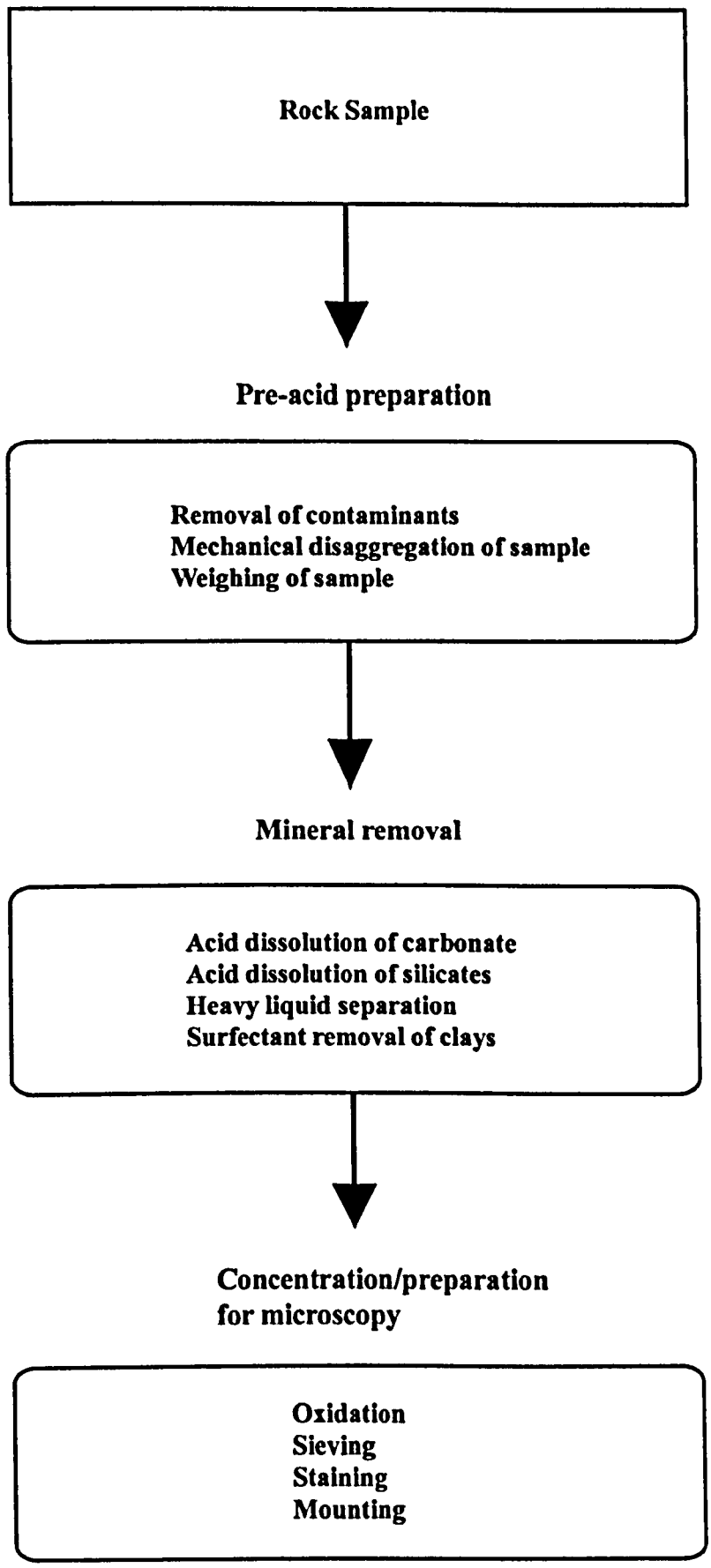


Fig. 2.6. Basic palynological processing. After Wood *et al.* (1996, fig. 1)

- This residue was accurately made up to 5 ml using polyvinyl alcohol to prevent clumping of the residue.
- Measured aliquots of 0.05 ml were withdrawn and made up as slides on round 13 ml glass coverslips, and then mounted using epoxy resin. This was repeated several times for each sample to allow the calculation of a reliable average
- Aliquots were counted for the total number of dinoflagellate cysts, miospores and acritarchs, and averaged to give *b*
- Cysts per gram of sediment then calculated for each sample using the formula:

$$\text{Cysts/ Gram} = (b \times 100/a) \times 2$$

$$= (\text{average aliquot count} \times 100/ \text{dry weight of sample}) \times 2$$

All samples for this study were prepared in the palynology laboratory of UCL by the author. All residues and slides will be stored at the British Antarctic Survey.

2.3 – DATA ANALYSIS

2.3.1 – Literature review of Dinoflagellate Cyst and Acritarch Palaeoenvironmental analysis

Understanding of cyst production in modern dinoflagellates and the fact that only 10% of the population produce decay-resistant cysts (Taylor, 1990), has led to the recognition that many, if not most, fossil preparations are unrepresentative of the original flora. Subsequently concerns developed about relating cyst distributions to those of the motile dinoflagellates. As a consequence a large amount of research has now been carried out into the conditions controlling the formation, distribution and

preservation of cysts (Dale, 1983). It was found that, despite only a small proportion of extant dinoflagellates encysting, those that do provide valuable evidence for the ecology of the motile stage of the life cycle. The conclusion reached by Dale (1983) was that the preserved cyst population *can* be used as a reliable proxy for changes in palaeoecology.

The assumption had long been made that cysts could be used as proxies, as studies from the 1970s had provided evidence pointing towards this link, e.g. Harland (1973), Davey & Rogers (1975), Wall *et al.* (1977). Following these early pieces of research, much effort was extended to developing methodologies that could be used in a consistent way to uncover relationships within the dinoflagellate cyst population that reflected the environmental conditions under which the dinoflagellates had lived. Advances in statistical analysis of the dinoflagellate cyst floras have led to the recognition that they can be grouped into ecologically significant associations.

There are now various methodologies for utilising the diverse and wide-ranging character of dinoflagellate cysts and other palynomorphs as proxies for palaeoenvironments. Many of these have been extensively reviewed in Davies *et al.* (1982). These are: -

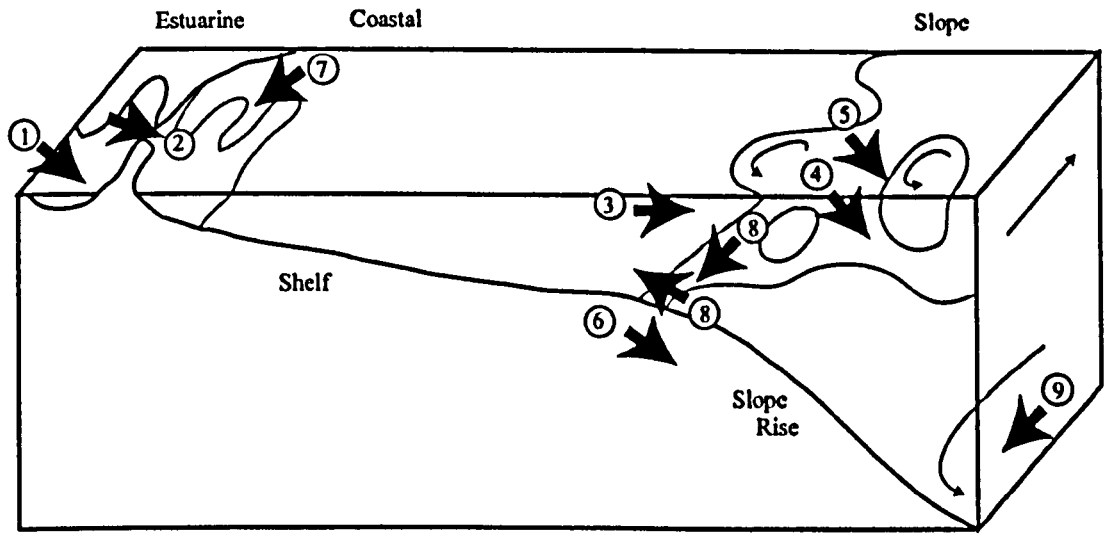
2.3.1.1 - Absolute abundance of dinoflagellate cysts

'The abundance of dinoflagellates can be measured as the number of dinoflagellates per unit mass or volume of sediment or water.' (Davies *et al.* 1982, p. 125). The majority of studies to determine absolute abundances of dinoflagellate cysts per unit of sediment have been carried out on Quaternary sediments (see review - Stover *et al.*, 1996). One of the most frequently cited works is Wall *et al.* (1977), a study that was carried out on modern marine sediments from differing settings

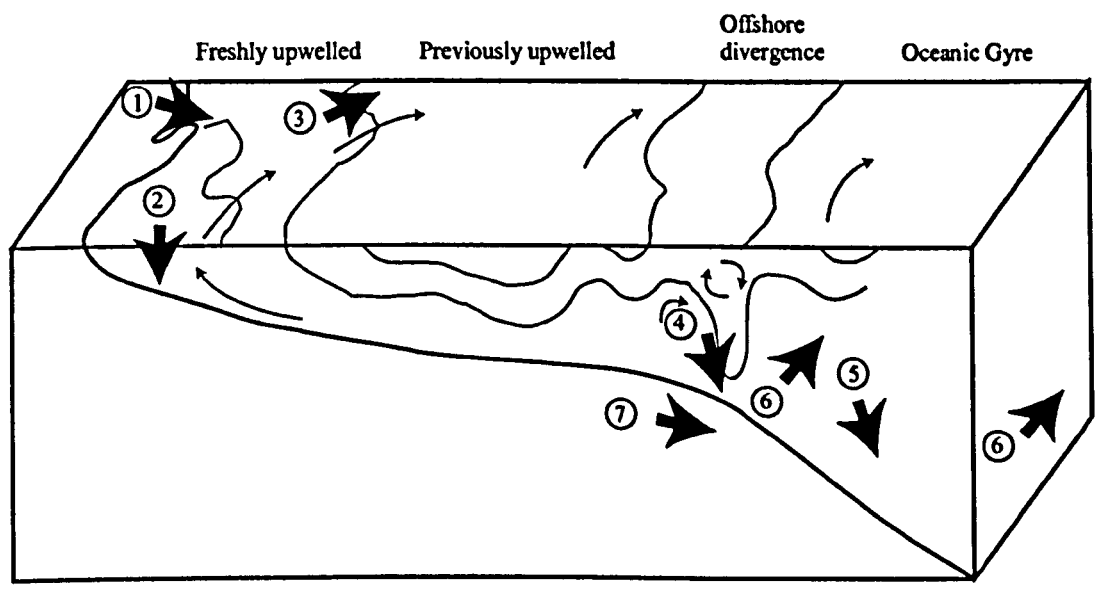
throughout the Atlantic Ocean and surrounding regions. After the analysis, two important distributional trends were identified using independent variables, with the implication being that the same trends could be recognised in fossil cyst data. These trends are: (1) changes in the assemblage composition in an inshore to offshore direction; (2) changes in the assemblage composition in a cross-latitudinal direction. The assemblage changes were of two kinds: a) the distribution of some species showing a restricted range related to inshore-offshore environments or latitude, and b) proportional changes in species composition also relating to inshore-offshore environments or latitude.

Wall *et al.* (1977) also studied factors that determined trends in cyst distribution and noted that 'Locations where important changes occur in assemblage composition in bottom sediments are closely coincident with hydrodynamic boundaries in the surface water circulation pattern which are usually marked by temperature-salinity discontinuities' (Wall *et al.* 1977, p. 166) (Fig. 4.1). Davey & Rogers (1975) and Wall *et al.* (1977) both found a correlation between the increasing concentration of dinoflagellate cysts and the decreasing concentration of terrigenous material in Quaternary sediments. In the former study, an increase was found in dinoflagellate cyst absolute abundances in a seaward direction; a further increase in abundance was also recorded offshore beyond the continental shelf edge.

Few absolute abundance studies have concentrated on pre-Quaternary sediments. However, Manum *et al.* (1989) looked at the variation of cyst occurrences in sediments of Eocene to Miocene age. It was found that changes in cyst abundance were apparently unrelated to the sedimentology of the sequence, i.e. abundance did not necessarily reflect sedimentological concentration, dilution or sorting.



A. 'SLOPE WATER' MODEL FOR CYST DISPERSAL



B. UPWELLING-OFFSHORE DIVERGENCE MODEL FOR CYST DISPERSAL

Fig. 2.7. Diagram to illustrate factors affecting cyst dispersal (after Wall *et al.* 1977, p. 167). Model A, 1= estuarine-lagoonal confinement; 2= outwash from lagoons; 3= offsetting coastal waters; 4= eddies of coastal water migrating seawards into offshore zones; 5= detached parcels of coastal water containing neritic microfloras; 6=winnowing of relict shelf deposits; 7= intrusion of coastal-shelf water into estuaries by mixing or bottom intrusions; 8= mixing, sinking and bottom influx of oceanic water with shelf water across the continental slope and the outer shelf zones; 9= currents of subpolar origin parallel to continental slope in intermediate waters. Model B, 1= localised inflows of estuarine water; 2= vertical sinking of the products of nearshore planktonblooms in upwelling areas; 3= seaward dispersal of nearshore bloom products by wind-driven transport; 4= vertical sinking of cells from the offshore zone of divergence; 5= vertical sinking from surface oceanic waters; 6= currents parallel to continental slope at intermediate depths; 7= winnowing of outer shelf and slope deposits during submarine erosion.

2.3.1.2 - Abundances of dinoflagellate cysts relative to other palynomorph groups

The use of relative abundances between different palynomorph groups is common in palaeoenvironmental reconstructions. They are frequently used to discern distance from shoreline of an assemblage by the relative influence of marine or terrestrial input. This can subsequently be related to transgressive-regressive events and relative sea level changes.

Among the more typical ratios used are spores and pollen: dinoflagellate cysts, or terrestrial palynomorphs: marine palynomorphs, (the marine component includes such groups as acritarchs and prasinophyte algae). Alternatively, these groups can be divided up to look for finer-scale changes. Working on the Early Jurassic of NW Germany, Prauss (1989) separated the three marine palynomorph groups (dinoflagellate cysts, acritarchs and prasinophytes) into different divisions and the resulting relative abundances were considered to reflect the palaeosalinity of the sequence.

2.3.1.3 - Dinoflagellate cyst species diversity and dominance

The use of abundances of particular species or morphotypes that can be placed into groups to indicate a specific palaeoenvironment, has become increasingly widespread. It has long been known that the distribution of marine palynomorphs, both stratigraphically and laterally, is environmentally controlled.

Wall (1965), in an acritarch study on the Lower Jurassic of southern Britain, showed that inner shelf assemblages were dominated by certain groups of species, e.g. *Micrhystridium* spp., while outer shelf and oceanic assemblages were dominated by different groups, the latter also being more diverse. The stratigraphical variation

between the two different assemblage types was discovered to be congruent with regressive-transgressive cycles of sedimentation during the Early Jurassic.

Dinoflagellate cysts have also been demonstrated to display similar patterns in species diversity and abundance (Stover *et al.*, 1996). During the early stages of a transgression and the regressive phase, e.g. relatively shallow water conditions, species diversity is low. In contrast, the later transgressive stages, e.g. deeper water, have much higher species diversity.

The recognition that the dominance of particular groups of species (often phylogenetically linked), as well as changes in species diversity, could be used as a palaeoenvironmental indicator has allowed this method to be applied to numerous Mesozoic and Cenozoic sections (Stover *et al.* 1996). A common way of applying this method to dinoflagellate cyst assemblages has been to use the gonyaulacacean ratio, i.e. the relative abundance of peridinialean species (predominantly near-shore, inner shelf species) to gonyaulacacean species (predominantly offshore, outer shelf species). Harland (1973) pioneered this approach, and it has been used to provide an indication of the distance to the shore.

Subsequent studies have produced evidence that other factors strongly influence the distribution of gonyaulacacean and peridinialean dinoflagellate cysts. Bujak (1984) found that high levels of nutrients, e.g. upwelling, were linked to high numbers of peridinialean dinoflagellate cysts, as did Powell *et al.* (1992). Versteegh (1994) used the gonyaulacacean ratio only as a productivity index, linking peridinialean dinoflagellate cysts to high-nutrient, i.e. eutrophic, waters and gonyaulacacean dinoflagellate cysts to nutrient-poor, i.e. oligotrophic, waters. In order to generate an index to indicate distance to shore, Versteegh used a ratio of terrestrial palynomorphs to marine palynomorphs (see 4.1.2).

In other work on dinoflagellate species diversity and abundance, Goodman (1979) produced a diversity index (total number of species present in a sample). Goodman also generated a dominance index. The correlation of these two indices, combined with the cluster analysis of the ten most abundant species in each sample, indicated that environmental factors (an inshore-offshore trend) were a major control over the distribution of the assemblages. Hudson (1990) carried out research into the effects of salinity on species distribution in the overall aquatic biota (not specifically marine palynomorphs) and found a consistent pattern. Maximum species diversity is reached in euryhaline (30-40‰) waters, diversity drops in polyhaline and mesohaline waters (5-30‰) and is lowest of all in brackish waters (0.5-5‰).

2.3.1.4 - Relative abundance of individual species

An individual species can, on occasion, be used as a palaeoenvironmental indicator. Previously, most of the work on these indicators has focused on the Mesozoic, although Cenozoic indicators have now been identified (Mudie 1986), and many of these are apparently climatically restricted (Stover *et al.* 1996).

2.3.2 - Marine Analysis Techniques

The principles of marine palynomorph distribution described above have led the workers who discovered them and others (Versteegh, 1994; Goodman, 1979) to develop simple statistical techniques to display the trend in question. The proxies chosen in this work have been used as they were felt by the author to best utilise the data, as well as most closely matching the already known environmental parameters of the Paleogene.

In order to simplify the analysis of the data generated during the present study, a number of curves have been established across the studied sections. Two of

these curves were generated *sensu* Versteegh (1994) and one *sensu* Goodman (1979). These are:-

(a) The Sporomorph/Dinoflagellate cyst curve, in which the terrestrial influence relative to the marine influence can be calculated by the equation:

$$S/D=nS/(nD+nS)$$

where n = number of specimens counted; S = Spores and pollen and D = Dinoflagellate cysts and acritarchs. A resulting ratio of 0.5-1 is indicative of being more proximal to the coast, while a ratio of 0-0.5 indicates a more distal position. The direction of the trend line is also important in establishing the regression or progression of the shoreline.

(b) The Peridinioid/Gonyaulacoid curve, in which changes in productivity, i.e. waters richer in nutrients, e.g. upwelling or estuaries, increase the numbers of heterotrophic peridinioid dinoflagellate cysts relative to the numbers of autotrophic gonyaulacoid dinoflagellate cysts that are tend to be more prevalent in nutrient-poor waters, e.g. outer shelf or oceanic waters. This can be calculated by the equation:

$$P/G=nP/(nP+nG)$$

where n = number of specimens counted; P = Peridinioid dinoflagellate cysts and G = Gonyaulacoid dioflagellate cysts. A ratio of 0.5-1 is indicative of higher productivity waters, while a ratio of 0-0.5 indicates less productive waters. It is again important to establish the direction of the trend line.

(c) Species Richness curve, here considered to be the mean average number of species present in the two preparations of each sample (after Goodman, 1979). It has been observed that dinoflagellate cyst richness changes in recognised patterns across an inner-shore to oceanic transect (see 4.1.3). Species richness

is a proxy that can be used, in conjunction with others, to ascertain changes in salinity, which in turn relates to changes in distance from shore, water depth and water mass character.

The relative abundance of individual species/genera is also plotted, as many species have specific environmental requirements and so their presence or absence can be a useful indicator of the environment of deposition. The species chosen for this are those previously recognised as forming monospecific assemblages and forming a major part of the flora through the section.

Apart from the quantitative analysis carried out as above, other methods of managing the data set were also used. Correspondence Analysis was applied to the data set to assess variations in assemblage composition with stratigraphic depth, with compositional change relating to environmental change.

Correspondence Analysis is a statistical analysis technique based on eigenvector routines with the data being standardised in a 'chi² matrix'. The pattern of numbers thus generated is analysed along eigenvectors (Davis, 1986) allowing the contribution a species makes to the total assemblage variation to be measured. The loading of these measurements upon each eigenvector can also be generated to allow stratigraphic analysis of assemblage trends (Nederbragt *et al.*, 2001).

Correspondence Analysis was run (by Dr S. Nederbragt) on a matrix of the 37 most abundant species in the López de Bertodano and Sobral formations together, and on the 31 most abundant species in the La Meseta Formation. Other selections of species were made to check for consistency. Scatterplots of loadings of species and samples on the first two correspondence axes were generated, which allows comparison of the clustering of sample loadings and the species responsible for these

results. Additionally, eigenvector loadings were also plotted against depth to permit analysis of stratigraphic variation.

3.2.3 – Literature Review of Spore and Pollen Palaeoenvironmental Analysis

Analysing the terrestrial microflora, e.g. spores and pollen, allows the reconstruction of the make-up of plant communities from the past (Tyson, 1995) as well as allowing the tracking of gross stratigraphic changes of these plant communities, which are thought to be primarily climate driven (Traverse, 1988). The use of terrestrial floras preserved in, and recovered from, a marine setting means that sedimentary factors and bias, e.g. transportation (Chaloner & Muir, 1968, MacPhail *et al.*, 1994), play a part in the final composition seen in the fossil record. However, changes seen in the assemblage will predominantly be due to actual changes in the plant community (Traverse, 1988).

The structure of the vegetational community present in the high latitudes of the Antarctic Peninsula area in the Early Tertiary would have been strongly dependent not only on temperature and precipitation, but also on light intensity. Previous reconstructions (e.g. Dettmann, 1994) have postulated widespread forests over West Antarctica. In order to support such a complex forest structure at latitudes higher than 60°S (Specht *et al.*, 1992), the canopy would have had to have been open, as the light intensity at the southern high latitudes would have been incapable of supporting an understorey in closed canopy conditions (Creber & Chaloner, 1987). Any canopy taxa present must have been relatively widely spaced (Dettmann, 1994).

Previous work (Dettmann, 1992, Askin, 1990) has shown that the most common elements of mid-high latitude Southern Hemisphere terrestrial microfloras during the Paleocene and Eocene are common Gondwanan groups, e.g. the

Podocarpaceae, Araucariaceae and Nothofagaceae, with smaller elements of other Gondwanan groups, e.g. the Proteaceae and Casuarinaceae. More geographically widespread background elements of the flora include the cryptogam spores, e.g. Cyatheaceae/Dicksoniaceae, Hymenophyllaceae/Osmundaceae, bryophytes and lycophytes.

All of the groups present in the Seymour Island flora have representatives still living in regions of the former Gondwana, e.g. Australasia, South America, or more widely spread. These, and similar species, have proved useful in developing models for the probable habitat, temperature, precipitation and latitude preferences of the different floral elements (Fig. 2.8). Previous climatic reconstructions for the West Antarctic region in the Paleocene have suggested a cool-temperate regime with temperatures of 11°C to 13°C (see Dettmann, 1992) and high annual precipitation. Previous studies of the terrestrial microfloras of Seymour Island (Askin, 1990) have also indicated a flora that is representative of a cool, temperate climate with high rainfall.

An important, and unique, component of Southern Hemisphere palaeobotany is the contribution made by the conifer family Podocarpaceae. The genus *Podocarpus* is widespread, growing in a number of ecological niches, and cannot in itself be used as a precise environmental indicator beyond a generalised 'cool temperate' climate. The common Seymour Island pollen *Phyllocladidites mawsonii* has been identified as belonging to the extant conifer species *Lagarostrobus franklinii* (Cookson, 1953). Today this grows only in the cool temperate conditions of western Tasmania, with rainfall in excess of 1200mm per year, i.e. rainforest (Askin, 1990). The pollen *Microcachryidites antarcticus*, linked

Miospore taxa	Suggested affinity	Distribution
BRYOPHYTIC/ LYCOPHYTIC SPORES		
Aequitriradites spp.		widely distributed, tropical and temperate. Moist conditions.
Stereisporites spp.	Sphagnaceae	
Foraminisporis spp.	Anthocerotaceae	
Retitriletes spp.	*Lycopodium (Lycopodia)*	Widely distributed, tropical and temperate. Moist conditions.
Ceratosporites spp. *Camarozonosporites* spp.	*Selaginella/ Lycopodium*	
Densoisporites/ Perotriletes spp	Selaginellaceae	
FILICEAN SPORES		
Cyathidites spp.	Cyatheaceae/Dicksoniaceae	Tropical & temperate, tree fern
Dictyophyllidites spp. *Triletes* spp.	Dicksoniaceae	Mainly tropical, tree fern
Gleicheniidites spp. *Clavifera* spp.	Gleicheniaceae	Tropical & subtropical regions, terrestrial fern
Baculatisporites spp. *Osmundacidites* spp.	Hymenophyllaceae/Osmundaceae	Tropical & temperate regions (hydrophilic/rainforest), filmy fern
Polypodiisporites speciosus (Harris) Kahn & Martin	Polypodiaceae	
Ischyosporites spp.	Schizaeaceae (cf. *Lygodium*)	Terrestrial fern
Azolla Lamarck	Salviniaceae	Aquatic conditions
Biretisporites spp. *Laevigatosporites ovatus* *Dictyophyllidites/Matonisporites* spp.	Multiple filicean source	
CYCADALES		
Cycadopites spp.	*Cycas*	

Miospore taxa	Suggested affinity	Distribution
GYMNOSPERMAE		
Araucariacites australis Cookson *Dilwynites* spp.	(Araucariaceae) *Araucaria*	Mainly Southern Hemisphere
Microachrydites antarcticus Cookson ex Couper	(Podocarpaceae) *Microachrys*	Tasmania, colder higher latitudes
Phyllocladidites mawsonii Cookson ex Couper	(Podocarpaceae) *Lagarostrobus/ Dacrydium*	Tasmania, rainforest tree
Podocarpidites spp.	(Podocarpaceae) *Podocarpus*	New Zealand, E.Australia, New Caledonia, New Guinea, E.Asia
ANGIOSPERMAE		
Nothofagidites senectus Dettmann & Playford	*Nothofagus* ancestral group	Extinct form
N. fusca- type	*N. fusca* group	S. America, New Zealand, Tasmania, cool temperate
N. menziesii- type	*N. menziesii* group	S. America, New Zealand, E. Australia, cool temperate
N. brassi- type	*N. brassii* group	New Caledonia, New Guinea, cloud forest
Beaupreaidites spp.	Proteaceae	Australia to E. Malaysia
Peninsulapollis spp.		N. Australia, S. America, tropical Pacific
Propylipollis spp.		Australia
Haloragacidites harrissii (Couper) Harris	Casuarinaceae	Queensland, New Guinea, New Caledonia. Lowland rainforest
Arecipites waitakiensis McIntyre	?Palmae	Tropical/sub-tropical, ranging into southern, cool temperate
Gemmamonocolpites pilulus Askin		
Clavamonocolpites polygonalis Askin	?Chloranthaceae	
Tricolpites spp	Multiple angiospermous source	
Liliacidites spp.		
Triporopollenites spp.	Unknown angiospermous source	

Fig. 2.8. After Dettmann & Thomson (1987) & Dettmann (1994)

to the conifer *Microcachrys*, occurs in colder, higher altitude settings (Askin, 1990). Assuming that there has been no significant shift in the environmental preferences of this group, the presence of podocarpaceous conifers indicates cool temperate conditions, the additional presence of *P. mawsonii*.suggests the existence of rainforest, while *M. antarcticus* suggests colder, higher altitude conditions

The patterns of distribution of *Nothofagidites* pollen throughout southern Gondwana and the history of their movement through the Tertiary has allowed the development of models linking the presence of some *Nothofagus* lineages to certain climate regimes and environmental preferences. The initial diversification of the *Nothofagus* lineages has been linked to Late Campanian-Maastrichtian tectonic and volcanic activity (Dettmann *et al.*, 1990).

Nothofagidites fusca-type is most frequently associated with cool temperate climates (Askin, 1990) while *Nothofagidites brassi*-type is linked to cooler temperatures (Pocknall, 1989). Extant forms of this lineage in New Caledonia and New Guinea grow at altitudes between 1500-3000m within the cloud forest zone (Askin, 1990), suggesting an additional preference for high precipitation. The *N. menziesii*-type is also associated with higher altitude settings and colder temperatures (Askin, 1990).

3.2.4 - Terrestrial Analysis Techniques

To reconstruct the terrestrial palaeoclimate and palaeoenvironment of the Seymour Island area during the Paleocene and Eocene, each family present will be plotted stratigraphically to show variation through the measured section. If a taxon is considered indicative of a specific environmental regime, e.g. *Lagarostrobus*,

then it will plotted individually. For further semi-quantitative environmental analysis of the terrestrial microflora present, selected taxa, both spores and pollen, will be placed into two ecological groups (McPhail *et al.*, 1994). These two groups, one reflecting plant community associations and the other on plant temperature regimes, are based on the recognition that: a) the relative abundances of spores and pollen reflect the distribution of plant communities, and b) that climate is the most important of all environmental factors determining plant distribution.

The ecological groupings of the spores and pollen present have been generated in two ways. In the first set, selected spores and pollen taxa have been placed in a broad plant community association.

1. Multistratal rainforest. The canopy components of this association are predominantly the podocarpaceous conifers, e.g. the taxa *Podocarpidites* spp., *Phyllocladidites mawsonii*, and *Microcachryidites antarcticus*. Other, smaller elements of the canopy may include the angiosperms represented by the pollen taxa *Proteacidites* spp. and *Nothofagidites* spp. The understorey component would probably also comprise elements of the Proteaceae and Nothofagaceae, as well as the tree fern group, *Cyathidites* spp. and *Dictyophyllidites* spp. The 'filmy fern' group, *Baculatisporites* spp. and *Osmundacidites* spp. have also been included in this association due to their role in ground cover, and also due to their role in growing epiphytically within the canopy.
2. Lowland rainforest. This community association comprises the angiosperm pollen *Haloragacidites harrisii*, which is linked to the Casuarinaceae genus *Gymnostoma* (Christophel, 1989). The Casuarinaceae are linked to lowland rainforest areas (MacPhail *et al.* 1994) in Northern Australia (Queensland) and

are considered to represent a separate community from the conifer-dominated rainforest above.

3. Open heathland/disturbed environment. This association is presented by the fern family Gleicheniaceae.

In the second set of ecological groupings, spores and pollen have been selected according to the recognised thermal regimes that their parent plants have been identified as living in. Nix (1982) generated an ecological classification relating plant growth to environmental variables (see MacPhail *et al.* 1994 for review). Three thermal groups were developed based on the mean annual air temperature (MAT): megathermal (>24°C), mesothermal (>14°C, <20°C) and microthermal (<12°C). These groups are also linked by intermediate categories. MacPhail *et al.* (1994) also described the linkage of the above categories with the widely used terms 'tropical', 'subtropical-warm temperate' and 'cool-cold temperate', respectively where moisture or light are not a limiting factor.

1. Mesothermal-megathermal temperature regime. Includes *Haloragacidites harrisii* (Casuarinaceae) and *Proteacidites* spp.
2. Microthermal-mesothermal temperature regime. Includes *Podocarpidites* spp., *Lagarostrobus/Dacrydiumites* spp. and *Nothofagidites* spp.
3. Microthermal temperature regime. Represented by *Microcachrydites antarcticus.*

3 - Chapter 3 - RESULTS

3.1 - OVERVIEW OF RESULTS

In total, 159 samples were processed and palynomorphs counted from Paleocene and Eocene sediments. Of these, 74 samples were from the Paleocene sediments of the López de Bertodano and Sobral formations, while 85 samples were from the Eocene sediments of the La Meseta Formation. The recovery of palynomorphs from the samples was generally good. Only 2 samples were barren and recovery was good to excellent in 108 of the samples, while palynomorphs were poorly represented in the remaining 47 samples. The preservation of the palynomorphs, where present, was very good and occasionally, excellent. The palynomorphs were without exception a pale yellow in colour, indicating no post-depositional heating or alteration.

3.2 - MARINE PALYNOMORPH COUNTS

3.2.1 - Paleocene counts

Thirty-three dinoflagellate cyst genera containing 40 species and 6 acritarch taxa were recovered. Absolute abundance count and relative percentage data are appended at the rear of the thesis in CD-ROM format. The absolute abundance of dinoflagellate cysts throughout section DJ 904 (Fig. 3.1) shows a low of 70.61/g sediment and a high of 11,963.96/g sediment. Four single abundance peaks are immediately obvious, these are sample numbers 904.44, 904.64, 904.48 and 904.92. Other abundance highs are seen in sample numbers 904.75, 904.79 and 904.81. The abundance lows are seen at sample numbers 904.61-63, 904.89 and 904. 104.

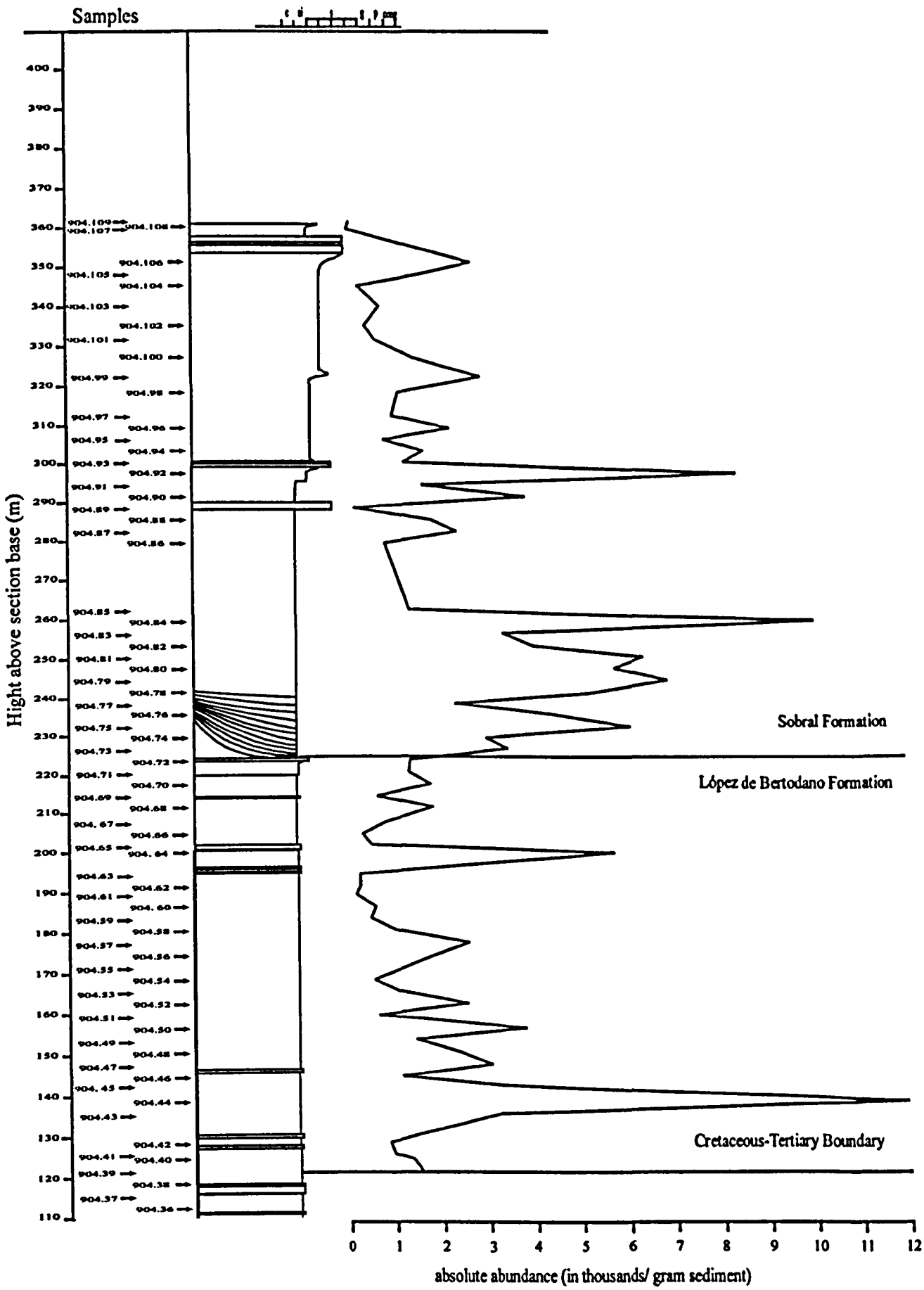


Fig. 3.1 Absolute abundance of Paleocene dinoflagellate cysts

The overall pattern of dinoflagellate cyst abundance shows a first peak of abundance at 904.44 (138m) (the flora is dominated by *Palaeoperidinium pyrophorum* - see Appendix 1) followed by a dramatic fall in abundance at sample 904.46 (143m). This is followed by a further gradual decline in abundance over the next 91 m (143-194m) which is ended by a second abundance peak at 904.64 (200m) (the flora here is dominated by *Trithyrodinium evittii* - see Appendix 1). Following this high, abundance falls abruptly once again after which it increases steadily over the next 54m (204-258m) to a third abundance peak at sample 904.84 (258m) (the flora is dominated by two species of *Spinidinium* - see Appendix 1). Abundance again falls abruptly between 258-278m, then rises over the next 19m (278-297m) to a fourth abundance peak at 904.92 (297m) (the flora is dominated by *P. pyrophorum* and *Spinidinium* spp. - see Appendix 1). Abundance then drops rapidly over the next 3m before showing an overall slower drop in abundance towards the top of the section (300-362m).

Acritarchs are the other component of the marine palynofloras, although they are greatly reduced in abundance since their acme in the Devonian (Strother, 1996). They vary in abundance from a low of 0/g sediment to a high of 5572.5/g sediment (Fig. 3.2). There is a small cluster of samples that form the largest abundance peak (samples 904.80-84) and two small clusters that form secondary abundance highs (samples 904.71-76 and 904.92-94). Both of these abundance clusters are dominated by the acanthomorph acritarch *Micrhystridium* Deflandre, 1937, emend. Sarjeant and Stancliffe, 1994 - see Appendix 1.

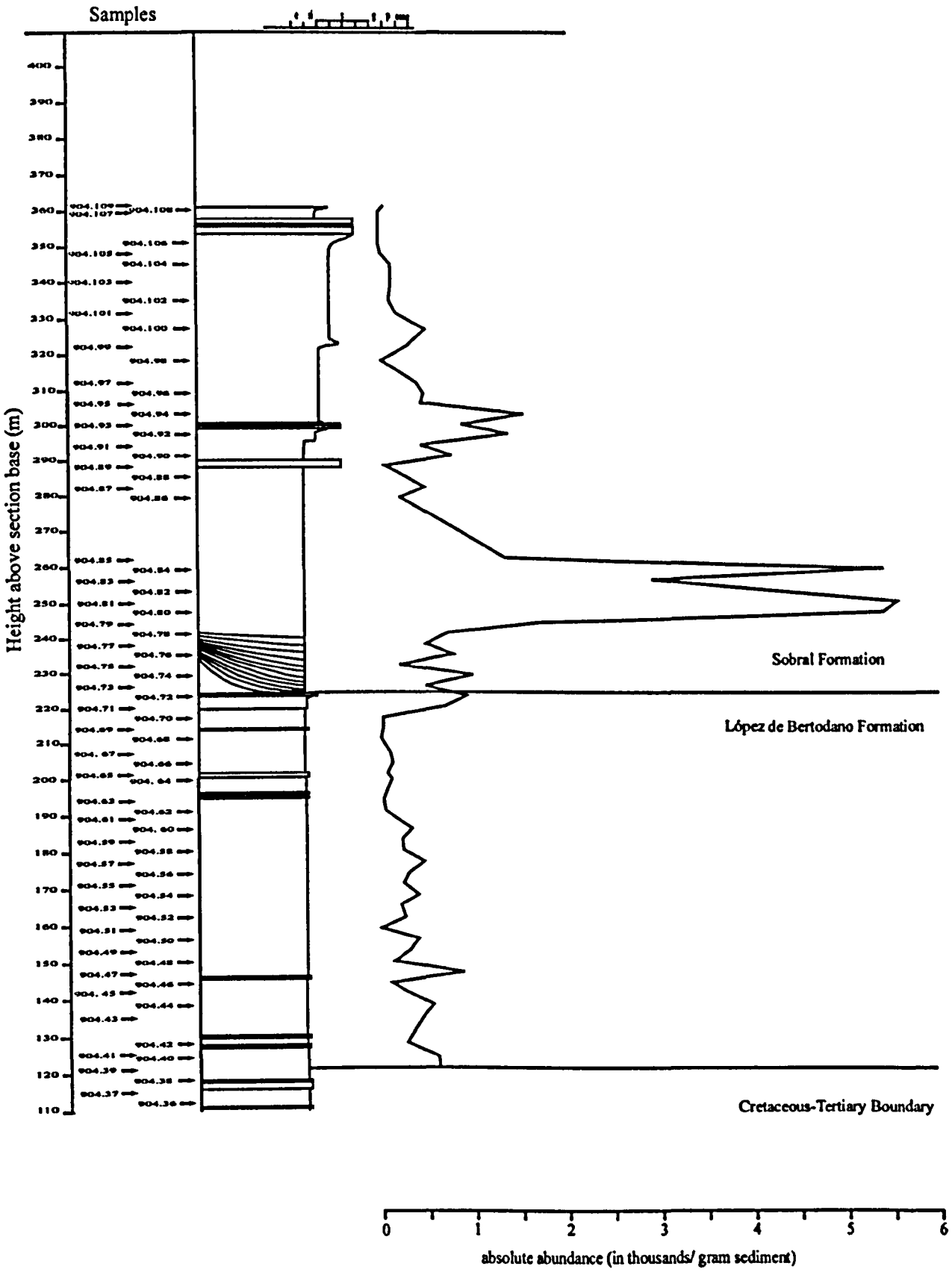


Fig. 3.2 Absolute abundance of Paleocene acritarchs

Acritarchs are present in relatively low numbers in the lowermost part of the section and decline slightly through the first 96 metres of the section (122-218m). Abundance then increases in the next 20 m in samples 904.71-76 (218-238m) to form the first abundance high of the section. Abundance then increases further in the next 21 m to form the second abundance high, in samples 904.80-84 (238-259 m) of the section. Acritarch abundance then tapers off over the next 29 m (259-288 m), at first steeply and then more gradually. Abundances rise gradually again over the next 14 m in samples 904.90-904.94, reaching a peak at 302 m. Following this final peak, abundances then decline over the next 60 m, towards the top of the section.

3.2.2 - Eocene counts

Twenty-eight dinoflagellate cyst genera containing 57 species (including 14 reworked species) and eight acritarch taxa were recovered (see Appendix 1). The dinoflagellate cysts vary in abundance from a low of 0/g sediment to a high of 36,445/g sediment (Fig. 3.3). Five abundance peaks are apparent and they occur in samples 801.1, 802.33, 802.23, 804.1 and 804.5. Abundance lows occur in samples 801.2-3, 802.13-1, 804.2, 804.9 and 804.11 and upward.

The lowermost part of section DJ 801 is in the upper part of the Sobral Formation and here abundance is relatively low. At the contact point between the Sobral and La Meseta formations, abundances increase dramatically in the first abundance peak in sample 801.1 (0 metres) before falling again to previous levels (abundance high due to domination by *P. pyrophorum* - see Appendix 1). Over the next 50m, abundances begin to increase, however, there is then a gap of 80 m (60-140m) in the section due to a lack of exposure at the outcrop (see Chapter 2). When the section is again exposed (at 150 m) abundances are lower and they stay low for

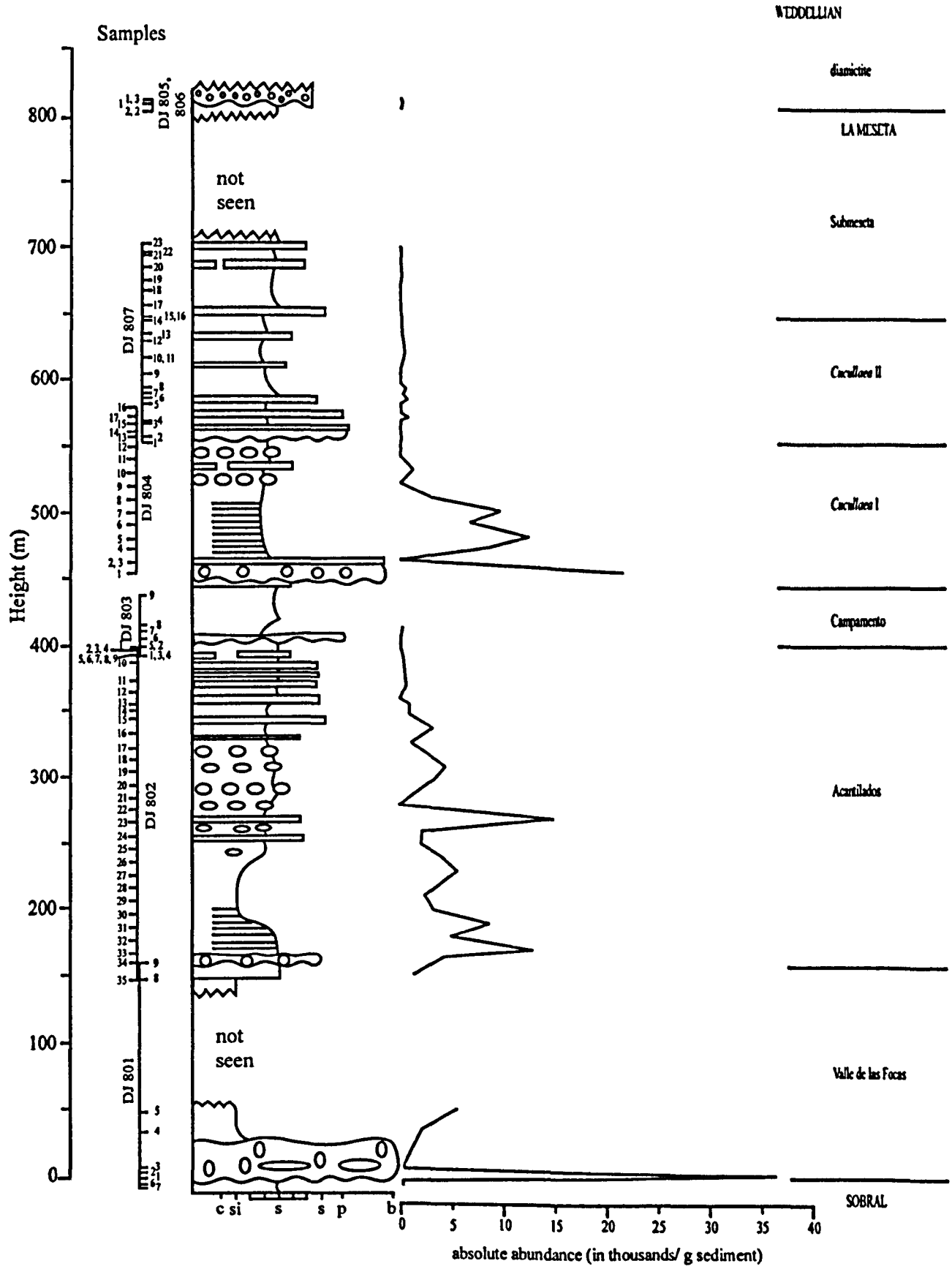


Fig. 3.3 Absolute abundance of Eocene dinoflagellate cysts

the next 15 m, after which there is a second single abundance peak in sample 802.33 (170m) (the flora is dominated by *Enneadocysta partridgei* - see Appendix 2). Following this, abundances decline for the next 95 m until there is a third abundance peak in sample 802.23 (at 275 m) (flora is again dominated by *E. partridgei*), abundances decline again over the next 38 m, from which point there is a sampling gap (420-455 m). Sampling resumes at 455 m and abundance is at a fourth peak in sample 804.1 (flora again dominated by *E. partridgei*). Abundance declines steeply from here, although it then rebounds to a fifth smaller and final abundance peak in sample 804.5 at 480 m (flora again dominated by *E. partridgei*). Following this final peak, abundance declines sharply for 40 m and then continues at a very low level for the next 305 m, until the top of the section, although there is a second gap in the outcrop exposure, with no collection possible for a 90 m section (715-805 m).

The acritarch component of the marine palynoflora varies in abundance from a low of 0/g sediment (Fig. 3.4) to a high of 23.854/g sediment. In the section there are two groups of samples showing an increased abundance, these are in samples 802.34-802.29 and 802.20-802.16. Both of these peaks are dominated by the acritarch *Enigmadinium cylindfloriferum* - see Appendix 1.

The absolute abundance of the acritarch flora is uniformly low for the first 50 m of the section. At this point there is then a sampling gap of 80 m (60-140 m). Just above the resumption of section and sampling, there is a 45 m (165-210 m) section of greatly increased acritarch abundance. Abundance reverts to its previous low levels for the following 55 m (220-275 m) after which there is a second section of increased acritarch abundance, although it is lower than the previous peak. This section lasts for

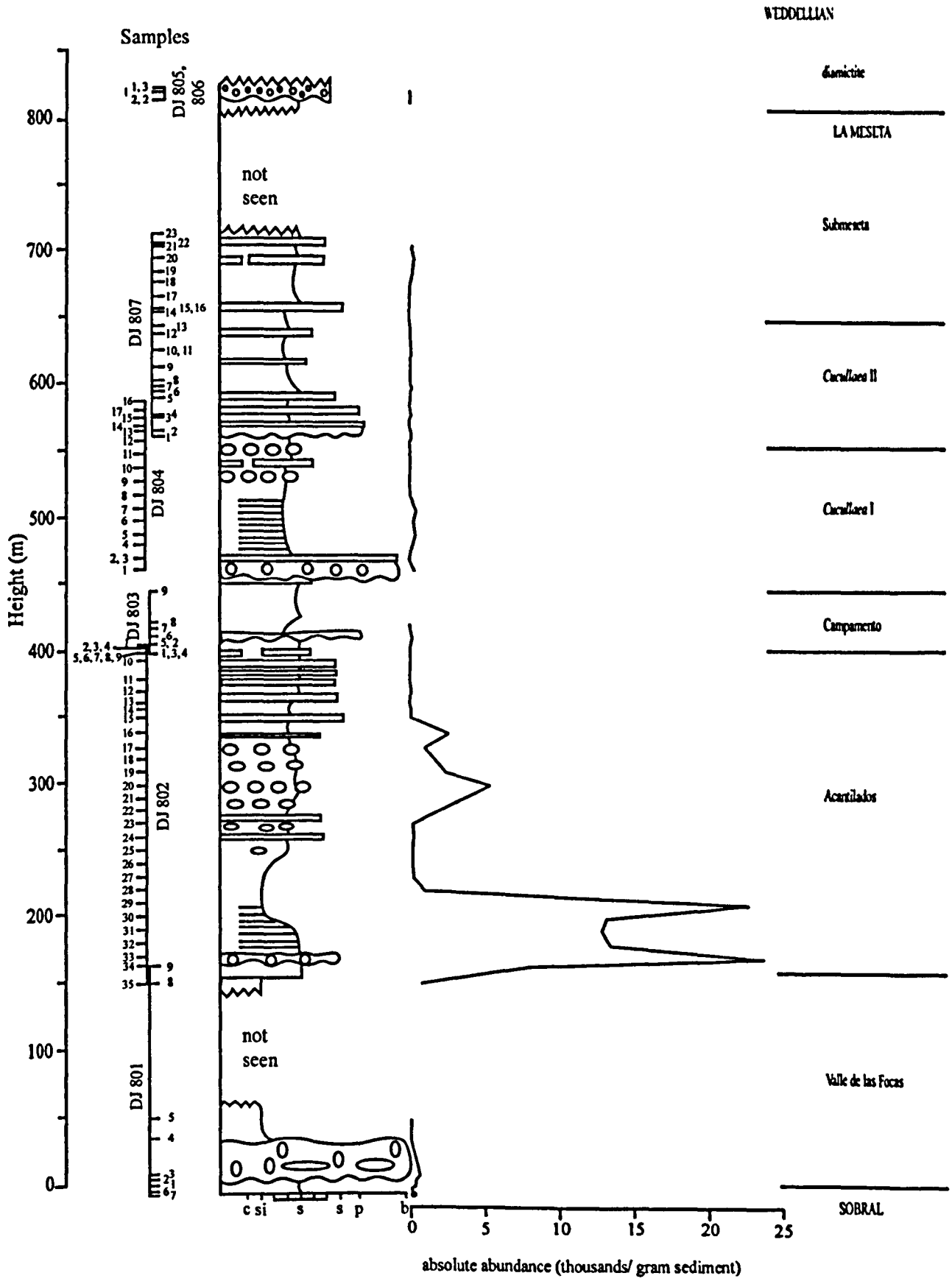


Fig. 3.4 Absolute abundance of Eocene acritarchs

45 m (295-240 m) and following this, abundance returns to low levels, remaining low for the upper 490 m of the section (340-830 m).

3.3 - TERRESTRIAL PALYNOMORPH COUNTS

3.3.1 - Paleocene counts

Forty-two genera and 57 species were recovered. Twenty-one genera are cryptogam spores, 6 are gymnosperm pollen and 15 are angiosperm pollen (see Appendix 2). The abundance of spores and pollen varies from a high of 11,815.99/g sediment to a low of 84.74/g sediment (fig. 3.5). In this section there are five abundance peaks in samples 904.40, 904.60, 904.69, 904.81 and 904.93. Four abundance lows are present in samples 904.51, 904.89, 904.104 and 904.107. The basal 120m of the measured section was not sampled (see Chapter 2).

The abundances of spores and pollen show a large amount of variation in the first 129 m of the section. Overall, the trend throughout this part of the section is from a first high in sample 904.40 (124 m) from where abundance declines through the next 59 m (124-183 m) including an abundance low in sample 904.51 (158 m). A second abundance high is present in sample 904.60 (186 m) although abundance then immediately returns to its previously low levels. Abundance then steadily increases over the next 25 m (189-214 m) to the third abundance peak in sample 904.69 (214 m). For the next 39 m (214-253 m) abundance remains generally high, although there is considerable variation through this part of the section. Following a fourth abundance peak in sample 904.81 (262m), abundance drops over the next 9 m

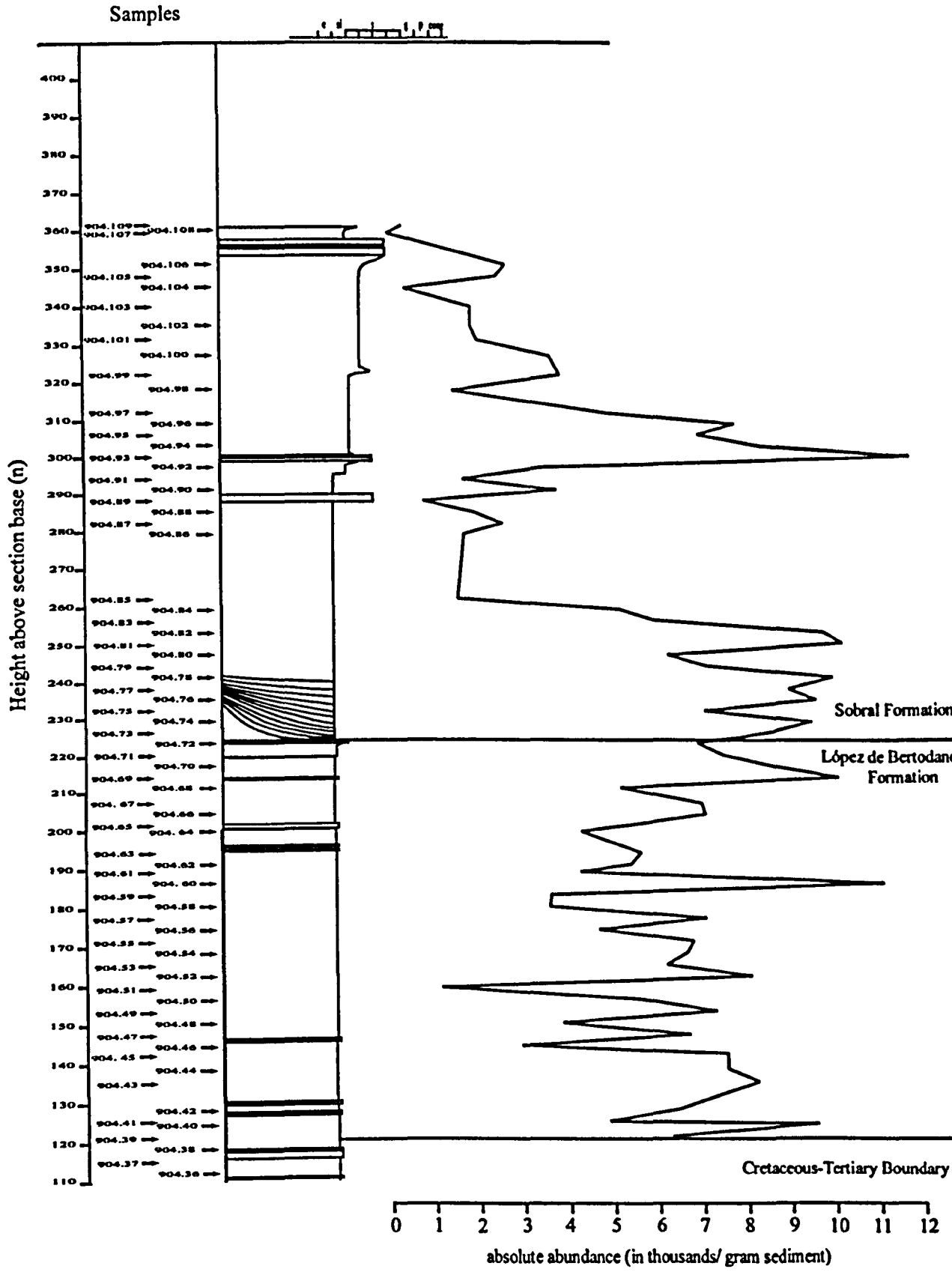


Fig. 3.5 Absolute abundance of Paleocene terrestrial palynomorphs

(253-262m) and it remains low for the next 32 m (262-294 m). Abundance then rises again to a fifth abundance peak in sample 904.93 (300 m), from which point it declines to the top of the measured section (300-362 m).

3.3.2 - Eocene counts

Thirty-seven genera and 50 species were recovered. Eighteen genera are cryptogam spores, 11 are gymnosperm pollen and 8 are angiosperm pollen (see Appendix 2). The abundance of spores and pollen varies from a high of 6542.050/g sediment to a low of 0/g sediment (fig. 3.6). In this section there are two groups of abundance peaks, these are in samples 802.33-27 and 804.4-804.8. Abundance lows are seen in samples 802.22, 802.13, 804.2, 804.9 and 804.11 upwards.

The abundance of spores and pollen in the basal part of the section is initially low, but it then rises to a small peak at 0 m before falling back to lower levels. Abundance then begins to rise over the next 40 m (10-50 m) before the outcrop becomes obscured and no samples can be recovered. When sample coverage resumes at 140 m, abundance is higher and for 65 m, through samples 802.33-27 (175-240 m) there is a first cluster of abundance peaks. Following these, abundance declines over the next 180 m (240-420 m). There is another sampling gap between 420-450 m. A second group of abundance peaks, in samples 804.4-804.8, begins at 470 m and continues for 50 m (470-520 m). From 520 m and upwards though the rest of the section (520-825 m), abundance declines, although there is some variation.

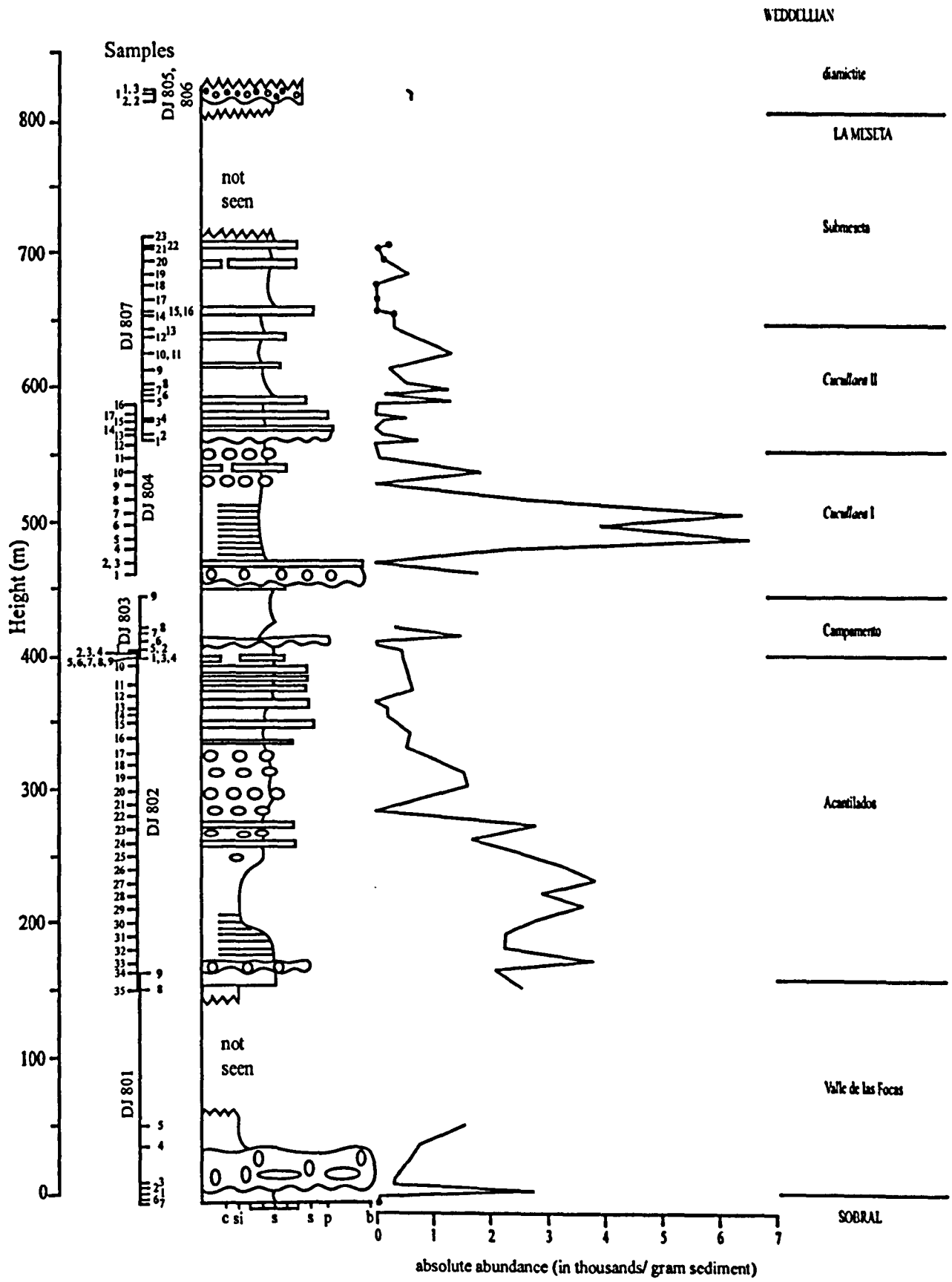


Fig. 3.6 Absolute abundance of Eocene terrestrial palynomorphs

3.4 - CORRESPONDENCE ANALYSIS

The data set generated (by Dr S. Nederbragt) for the marine floras of the Paleocene and Eocene is given in Appendix 3

Chapter 4 – Environmental interpretation and analysis of marine floras

4.1 - Palaeoenvironmental Analysis

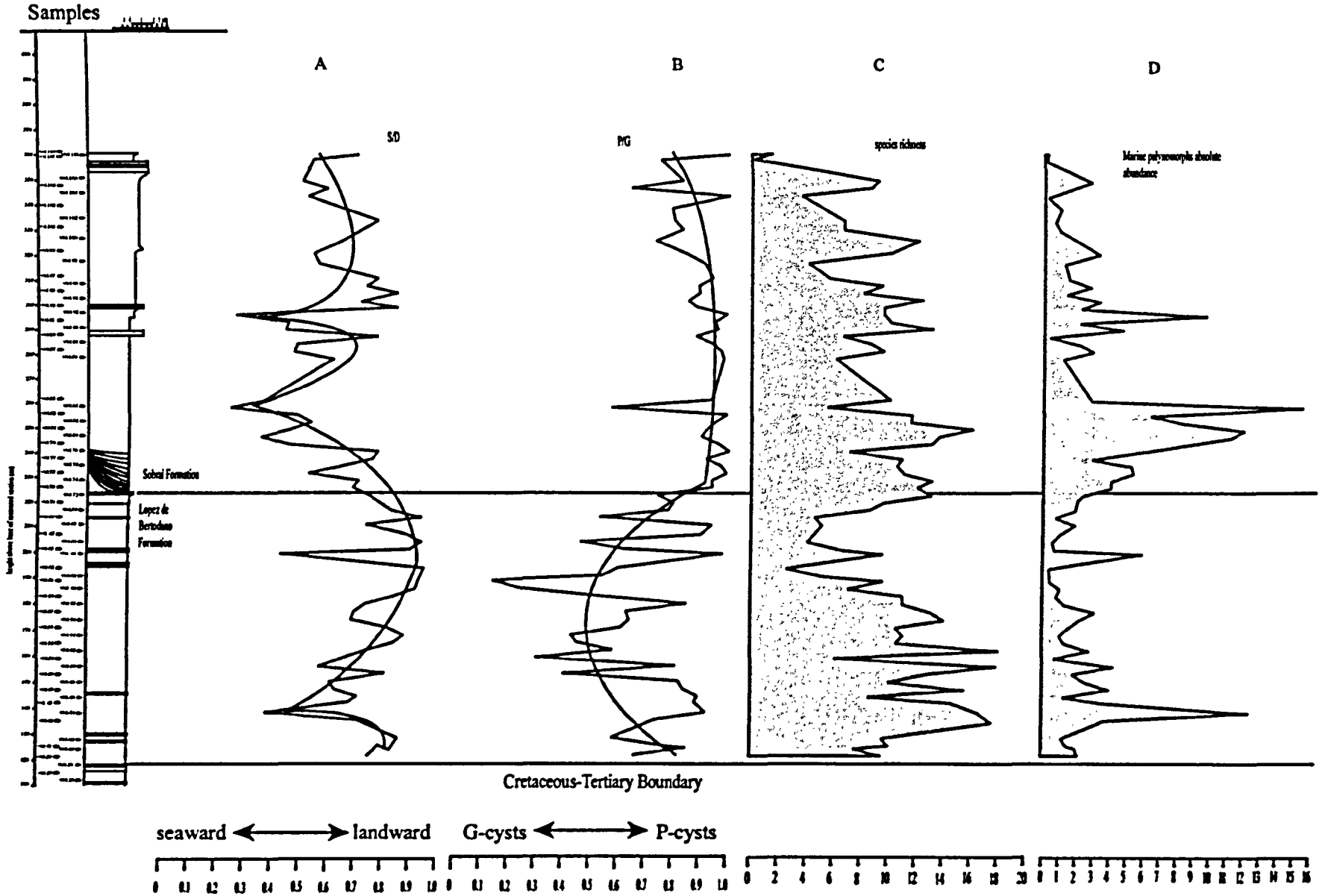
4.1.1 – Quantitative Analysis

4.1.1.1 - López de Bertodano Formation

For the interval studied, immediately above the K-T Boundary, the quantitative data produced shows distinct, consistent trends in each of the four palaeoenvironmental proxies generated from the data set.

In Fig. 4.1A, the plot generated for the spore and pollen: dinoflagellate cyst and acritarch ratio (S/D ratio) that is used as an indicator of the distance from the shoreline, shows the most significant trend to be an overall increase in the S/D ratio through most of the measured thickness (DJ 904.44-63). This represents a shift from a more marine-influenced flora to a more terrestrially-influenced flora. Other features of the S/D curve in this section are the strongly terrestrially-influenced basal four samples (DJ 904.39-42) and the decline in terrestrial influence apparent in the uppermost samples of the formation (DJ 904.64-72). There is also peak of strong marine influence (DJ 904.64) towards the top of the increased terrestrial-influence trend.

The analysis indicates an overall regressive trend occurring through the majority of the López de Bertodano Formation that is present. The lowermost samples mentioned above indicate an initial terrestrially-dominated flora,



4.1. A. Plot of S/D ratio, B. Plot of P/G ratio, C. Plot of species richness, D. Plot of absolute abundance of marine palynomorphs

close to the shoreline. This was prior to a distinct transitional shift to a strongly marine- dominated assemblage (DJ 904.43-4), suggestive of a transgression caused by a rise in relative sea level. From this marine-dominated assemblage (DJ 904.44) the shallowing-up, regressive trend, which dominates the upper part of the formation (with the exception of the marine-influenced peak at DJ 904.64), begins.

The uppermost López de Bertodano Formation exhibits a decline in S/D ratio, which is indicative of an increase in marine influence, although the flora is still strongly terrestrially-dominated. Overall the S/D ratio for the López de Bertodano Formation assemblage is indicative of a strong terrestrial influence, suggesting a depositional environment relatively close to the shoreline. This interpretation is supported by Macellari (1988) who interpreted the earliest Paleocene sediments (Unit 10 of Macellari, 1988) of the upper López de Bertodano Formation as being an inner shelf environment, deposited during a regressive phase (Fig. 4.2 & 4.3).

Fig. 4.1B, which displays the peridinialean: gonyaulacalean ratio (P/G ratio), shows an overall trend of decrease, then a corresponding increase in the productivity proxy. The P/G ratio also shows a high amount of variation, which can obscure the overall trend. The decline in the curve begins in DJ 904.44 and reaches its low point in DJ 904.61, from where it increases to the top of the formation. The exception to this trend is seen in the lowermost four samples (DJ 904.39-42) that have a markedly lower P/G ratio value than samples subsequent to them.

The P/G data can be interpreted overall of being dominated by peridinioid dinoflagellate cysts and thus of being indicative of high-nutrient, productive water, i.e. eutrophic conditions. The overall trend of decline and rise of the water productivity could be linked either to a change in the levels of nutrient input or to a

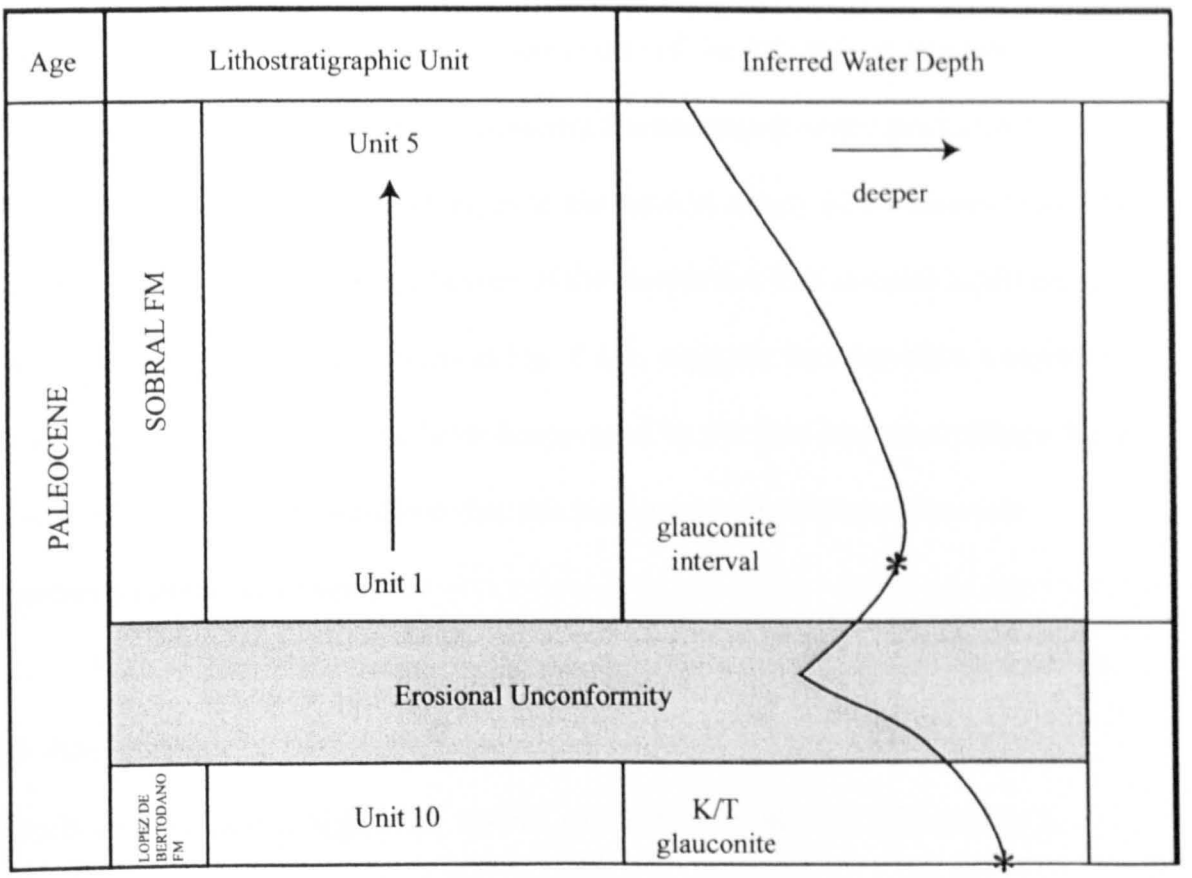


Fig. 4.2. The interpretation of depositional sequences of the uppermost López de Bertodano and Sobral Formations, after Macellari (1988, fig. 18)

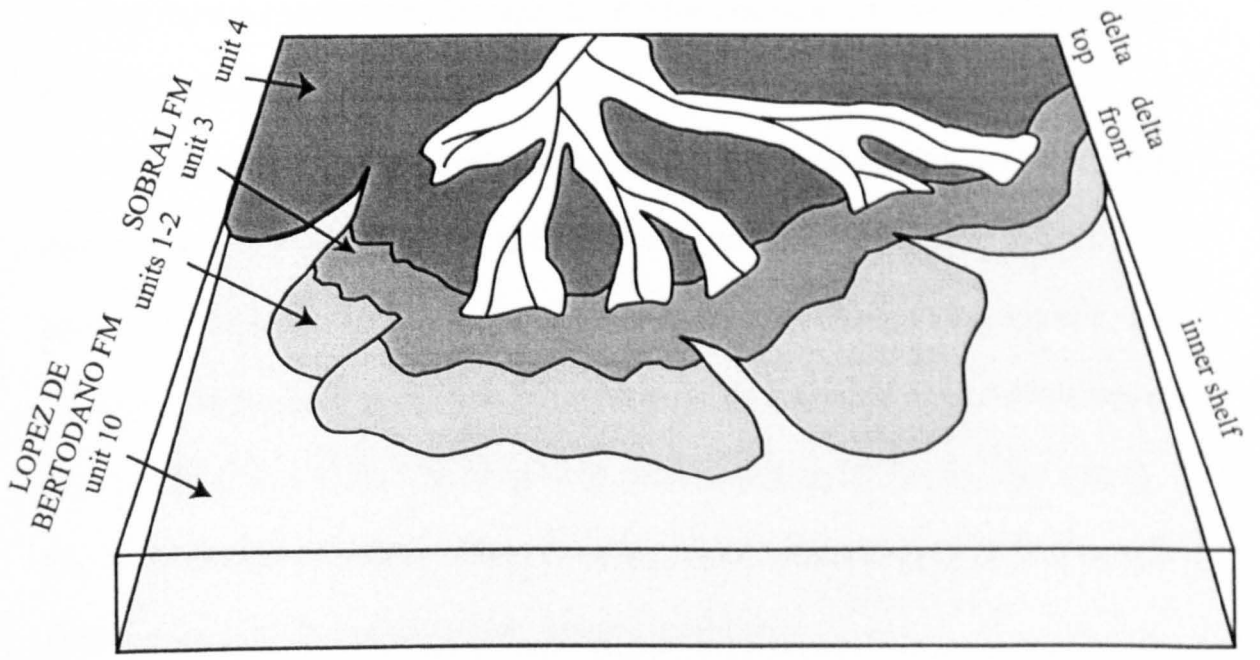


Fig. 4.3. Depositional environments of the López de Bertodano and Sobral Formations, after Macellari (1988, fig. 19).

change in relative sea level causing the preservation of a more distal flora from more oligotrophic, nutrient-poor waters. The variability of the P/G ratio throughout the section suggests short-lived, rapidly changing fluctuations in water productivity, which is also possibly linked to changes in the nutrient supply in the waters where the assemblage was living. The significance of the lowermost four samples again being distinct from the rest of the flora, as in Fig. 4.1.A, suggests that they form a separate flora, reacting to conditions that have disappeared by the time later assemblages have developed. It is a strong likelihood that this lowermost assemblage represents recovery from the K-T event.

The species richness curve is displayed in Fig. 4.1C. The major trend within this data set is similar to that of the previous two proxies, e.g. a decline is seen in species richness from a high in DJ 904.43, to a low in DJ 904.63, followed by a further increase so that species richness at the top of the formation is similar to that at the base of the trend. The basal four samples again appear as a distinct cluster, separate from the major trend through this section and an anomalous species richness high is seen in DJ 904.64.

This trend can be interpreted as a probable decline in relative sea level, with a change from a low energy setting to higher energy conditions, possibly above the storm wave base. The species richness low is reached in 904.63. Following this, the increase to higher species richness levels seems to be indicative of a return to more normal marine conditions. The anomalous richness high in DJ 904.64, also seen in Fig. 4.1A, indicates a sudden influx of marine, highly productive, conditions during otherwise apparently shallow water, stressed conditions.

Fig. 4.1D shows the absolute abundance data of marine palynomorphs through the upper López de Bertodano Formation. It shows an overall trend similar to

those described above. The most striking feature of the abundance curve is a high in DJ 904.44, from which point abundance declines. The low point of the curve is seen in DJ 904.63. Abundance increases again through to the top of the formation. An anomalous abundance peak is seen in DJ 904.64 (see above). The basal four samples in the section (DJ 904. 39-42), are again seen to be a distinct cluster. As in Fig. 4.1A, DJ 904.43 appears as a transitional assemblage between the basal cluster and DJ 904.44.

The absolute abundance data can be interpreted to show the presence of a low abundance flora in the lowermost part of the section, probably due to low relative sea level or stressed conditions, e.g. a K-T recovery flora. There is then a significant increase in abundance as relative sea level increased. This is followed by a longer-term fall in relative sea level, seen by the decline in abundance. From an abundance low in DJ 904.63, which may reflect a relative sea level low, abundance increases though the upper part of the section. DJ 904.44 again is an anomalous abundance peak in this increasing abundance trend in the uppermost López de Bertodano Formation.

The relative abundance of individual species shows other trends through the López de Bertodano Formation. The most apparent is the overall shift in dominance of the assemblage by *P. pyrophorum* and *Spinidinium* spp. in the lower part of Unit 10 (Fig. 4.3), to the dominance of the assemblage by *T. evittii* in the upper part of Unit 10. *P. pyrophorum* is a peridinioid dinoflagellate cyst (P-cyst) and an opportunistic species known to be very common in high-nutrient waters e.g. upwelling or estuarine waters, where it can form algal blooms.

Helby *et al.* (1987, p. 69) noted that as a 'significant accessory form, *P. pyrophorum* was often prominent' in the Early Danian *T. evittii* Acme Zone and that

both species are present in high abundance in this sequence, however, they do appear to be mutually exclusive in the López de Bertodano Formation. (Fig. 4.5). This is also the case in certain contemporaneous sections from New Zealand (Willumsen, 2000) and was theorised as being due to the inability of *T. evittii* to survive in colder, nutrient-rich upwelling waters where *P. pyrophorum* can thrive.

Trithyrodinium evittii (also a P-cyst) has been identified as a warm water species that originated in the low latitudes by Brinkhuis *et al.* (1998), who recognised that it migrated to both the northern and southern hemispheres apparently simultaneously immediately after the K-T Boundary event. Brinkhuis *et al.* (1998) have attributed this to a global warming event in the Early Danian. The dominance of the flora that *T. evittii* displays in this study (a high of 80.04%) suggests that this warming was a distinct event that in the Seymour Island area may have affected water temperatures to such an extent that it displaced *P. pyrophorum* from the assemblage and also affected the distribution of *Spinidinium* spp. (Fig. 4.5).

4.1.1.2 - Sobral Formation

The Sobral Formation, which immediately overlies the López de Bertodano Formation displays a wider variety of trends in the curves generated by the proxies used than the preceding formation.

Fig. 4.1A, the S/D ratio curve, the trend through the Sobral Formation is for a decline through the lower part of the Sobral Formation (DJ 904.73-84) from initially high levels at the López de Bertodano Formation and Sobral Formation contact. The S/D ratio then increases again (DJ 904.85-9) to levels comparable to those at the base of the Sobral Formation. Samples DJ 904.90-2 then display a sudden and short-lived excursion to lower S/D ratio levels. Following this, S/D ratio levels increase to high

levels (DJ 904.93) and for the rest of the Sobral Formation section (DJ 904.94-109), levels remain high, although there is a slight decline in values.

The interpretation of the Sobral S/D values indicates a continuation in the lowermost Sobral Formation of the increase in marine-influence seen in the uppermost López de Bertodano Formation. The rise in relative sea level seen in the lower Sobral Formation (DJ 904.73-84) suggests the occurrence of a transgression that may have begun in the uppermost samples of the López de Bertodano Formation and continued across the contact. The congruity of the curve across the boundary suggests that the erosional unconformity that defines the surface in some parts of Seymour Island does not represent a significant length of time. This indicates that the increase in relative sea level is indeed a continuation of the same trend. Askin (1988a) considered the lower unit of the Sobral Formation (Tpsl of Sadler, 1988) to represent a continuation of an overall regressive trend initiated during deposition of the López de Bertodano, while Macellari (1988) considered it to represent a 'new depositional cycle' (Macellari, 1988 p. 50). He interpreted the depositional environment as being part of a prograding delta fan in a relatively shallow, low energy setting (Figs 4.3 & 4.4). In Fig. 4.1A, maximum relative sea level is reached at DJ 904.84, which shows strong marine-influence. From this point, relative sea level is observed to fall, as samples (DJ 904.85-89) show an increasing terrestrially-influenced signal. This reaches a peak in DJ 904.89, when relative sea level is at its lowest. The sudden and short-lived excursion to low S/D ratio levels described above (DJ 904.90-2) can be attributed to as an increase in relative sea level and a short-term return to marine conditions. This event terminates abruptly (DJ 904.93) as relative sea level drops sharply and the flora becomes strongly terrestrially-influenced. From DJ 904.94 upwards through the rest of the section, S/D ratios remains high. This is indicative of

low relative sea level and the floras are characterised by high levels of terrestrial input. There is a slight decline in relative sea level through this part of the section, with some variation, although no cyclicity is apparent as seen in the lower Sobral Formation.

The productivity index (P/G) for the Sobral Formation, Fig. 4.1B, shows no major changes through the measured section. The assemblage seems to be characterised by an abundance of heterotrophic peridinioid dinoflagellate cysts, producing a very high P/G ratio. Features that are apparent on the P/G curve include a peak of lower P/G value (DJ 904.84) with a higher proportion of autotrophic gonyaulacoid dinoflagellate cysts. The other apparent feature is a slight decline in P/G ratio values (DJ 904.94-109) with some increased variation in results.

The P/G ratio can be interpreted to show consistently high water productivity, e.g. eutrophic conditions, in the Sobral Formation. The overall lack of variation indicates that nutrient levels remain consistently high, either as a result of terrestrial input or offshore upwelling. The consistency of the P/G ratio, despite the relative sea level changes seen in Fig. 4.1A, would suggest the former. The presence of 'restricted environments...at several intervals' was described by Askin (1988a, p. 141) in her lower zone 5 as a consequence of her discovery of near monospecific assemblages. This was explained as representing peridinioid dinoflagellate cyst blooms in restricted environments or conditions.

The species richness curve, Fig. 4.1C, shows less clear cut trends than were seen in the López de Bertodano Formation. The overall trend through the Sobral Formation is of decline. Despite this, considerable variations are seen and diversity does remain high in parts of the section. Species richness remains constant through the contact between the López de Bertodano and Sobral formations and then

increases to a richness peak (DJ 904.81). A single species richness low (DJ 904.78) is apparent during this otherwise increase in richness. From the peak (DJ 904.81), species richness declines through the rest of the section. There are several peaks of higher species richness within this overall trend (DJ 904.90, 904.90, 904.99-100 & 904.104-5).

The species richness curve in the Sobral Formation can be interpreted to show a continuation of the increase in relative water depth across the contact between the López de Bertodano and Sobral formations. This increase in water depth appears to continue until maximum species richness is reached in DJ 904.81. The anomalous species richness low (DJ 904.78), indicative of a sudden fall in relative sea level, correlates with a peak of increased terrestrial influence (Fig. 4.1A). The decline in species richness seen through the rest of the section can be interpreted as a decline in relative sea level, changing palaeoenvironmental conditions to those unsuitable for most dinoflagellate cyst species. The peaks of elevated species richness mentioned above (DJ 904.90, 904.90, 904.99-100 & 904.104-5) also correlate with areas of increased marine influence (Fig. 4.1A). This is evidence of short-lived fluctuations in sea level affecting both the S/D ratio and species richness, however the lack of any effect on the P/G ratio suggests that the nutrient source causing the presence of high levels of peridinioid dinoflagellate cysts is independent of relative sea level changes and is related instead to terrestrial input.

The absolute abundance curve (Fig. 4.1D) shows an overall trend similar to that seen in the species richness curve (Fig. 4.1C), e.g. one of overall decline, but with some samples showing high abundance. Absolute abundance increases across the contact between the López de Bertodano and Sobral formations, and continues to increase to a peak at DJ 904.84. Two small peaks of lower abundance are seen within

this larger increase (DJ 904.77 & 904.83). Following the abundance high, absolute abundance declines to a low (DJ 904.89). There then follows a short interval of marked absolute abundance increase (DJ 904.90-92). Subsequent samples through to the top of the section show a steep decline in abundance, although two minor abundance peaks are seen (DJ 904.99 & 904.106).

The interpretation of the absolute abundance data would indicate that the erosional unconformity separating the two formations represents a negligible length of time, given the lack of any distinguishable break in the pattern of marine palynomorph abundance. This corresponds to the observation made above that the localised unconformity is not temporally significant. The absolute abundance data indicates an increase in relative water depth in the lower Sobral, reaching a maximum depth in DJ 904.84. The two anomalous peaks of lower abundance noted above (DJ 904.77 & 904.83) correlate with two peaks of increased terrestrial abundance (Fig. 4.1A) and decreased species richness (Fig. 4.1B). All three proxies indicate short-lived falls in relative water depth at these points in the section as the assemblages become more characteristic of nearer the shoreline. Following the relative sea level high (DJ 904.84), relative sea levels first decline as a regression occurs (DJ 904.85-89) and then relative sea levels rise during a transgressive period (DJ 904.90-92) in what appears to be a small regressive-transgressive cycle. The decline in absolute abundance in the upper part of the Sobral Formation seems to be clear evidence of a fall in relative sea level, albeit with two brief peaks of elevated relative sea level corresponding to those in Fig. 4.1A (DJ 904.99 & 904.106).

The abundance of individual species through the Sobral Formation shows a distinct change from the pattern of distribution in the López de Bertodano Formation

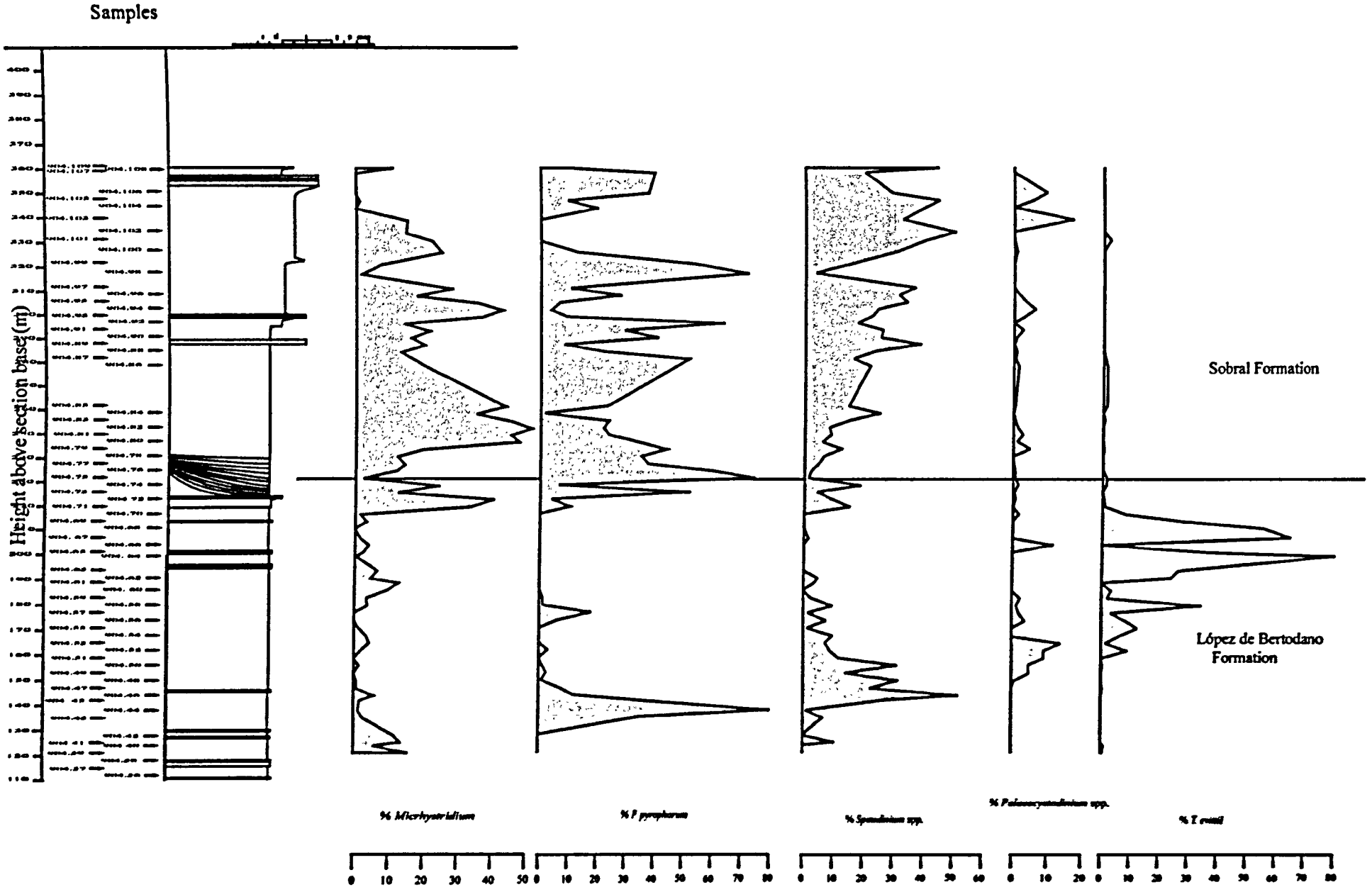


Fig. 4.4. Diagram of dominant marine palynomorph species of the Lopez de Bertodano and Sobral formations (as rel. %).

(Fig. 4.4). *Trithyrodinium evittii* is absent from the assemblage in any significant numbers, while the assemblage is instead dominated by the dinoflagellate cysts *P. pyrophorum* and *Spinidinium* spp. and the acritarch *Micrhystridium*. *Micrhystridium* and *P. pyrophorum* show a mutually exclusive relationship through the sampled Sobral section, while *Spinidinium* spp. show a steady increase in abundance through the section.

The implications for the distribution of *Spinidinium* spp. is currently unclear, while *Micrhystridium* is known to be more numerous in near-shore environments (Wall, 1965) and *P. pyrophorum* favours high-nutrient waters (see above).

4.1.1.3 - Cross Valley Formation

The Cross Valley spot samples lack evidence for marine floras, except for the presence of one single unidentifiable peridinioid dinoflagellate cyst in sample DJ 908. It has therefore proved impossible to provide any evidence for the palaeoenvironment or age of the Cross Valley Formation from the marine palynofloras of these samples.

4.1.1.4 - La Meseta Formation

The style of both sedimentation and palynoflora distribution in the La Meseta are very different to those seen in the Palaeocene sediments described above. The S/D ratio (Fig. 4.5A) shows a complex cyclical distribution through the La Meseta Formation. The upper La Meseta Formation (*Cucullaea* I, *Cucullaea* II and Submeseta members) shows an overall upwards trend in the S/D ratio, whereas this is not apparent in the lower part of the formation (Valle de las Focas, Acantilados and Campamento members) even though there is considerable variation within each member.

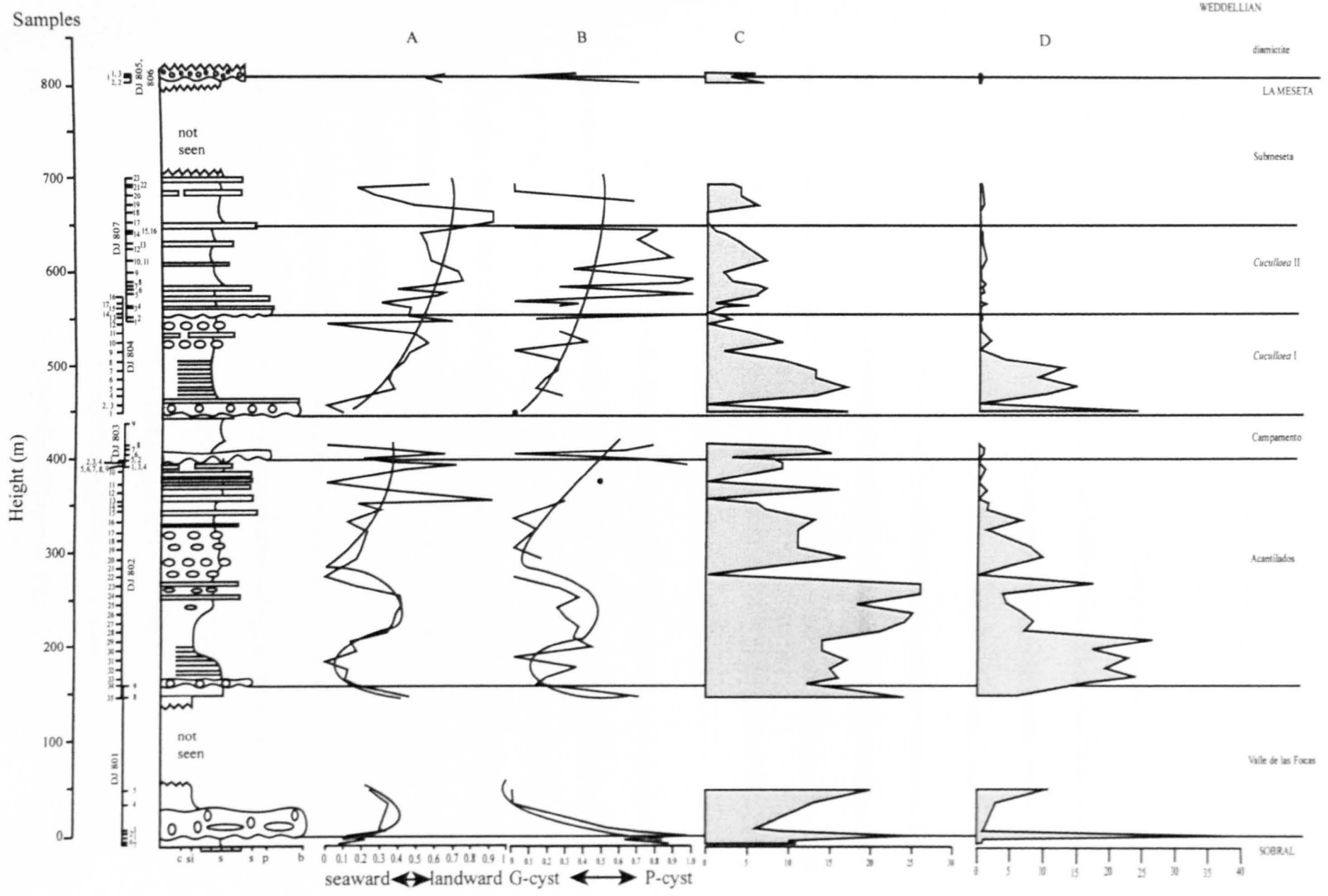


Fig. 4.5. A.Plot of S/D ratio, B. Plot of P/G ratio, C. Plot of species richness, D. Plot of absolute abundance of marine palynomorphs

A large thickness of the basal La Meseta Formation, the Valle de las Focas Member, was not sampled, fluctuations are apparent within this member, but insufficient data are available to allow a reliable analysis. However, the section that is present does show low S/D ratio values (DJ 801.2-5) and the uppermost sediments (DJ 801.8) show a decrease in value across the Valle de las Focas and Acantilados boundary. Reduction of the S/D ratio continues into the Acantilados Member (DJ 801.9/802.34-14) where floras remain marine-influenced, although the lower Acantilados Member is also characterised by a broad-based positive excursion (DJ 802.29-24) of less marine-dominated palynofloras. The assemblages are still overall marine-influenced. This dominance continues until the upper part of the member (DJ 802.13-803.4) where terrestrial influence increases and a large amount of variation is seen between strongly marine- and terrestrially-influenced floras.

The trend of fluctuations between marine- and terrestrial-dominated floras continues in the Campamento Member. The lowermost sample in this member (DJ 803.5,2) is strongly marine–influenced, with later samples showing the fluctuations previously noted. The Campamento Member, as with the Valle de las Focas Member, has not been sufficiently sampled to allow an overall trend to become apparent, or reliable analysis to be made.

The *Cucullaea* I Member shows a more distinct trend with fewer internal fluctuations. The S/D ratio indicates a change in assemblage in the lowermost samples from a strongly marine-influenced setting (DJ 804.1-8) to a more terrestrially-influenced, near-shore setting (DJ 804.9-14, 807.1-2). The increase in terrestrial influence continues through the uppermost *Cucullaea* I Member and into the *Cucullaea* II Member. The palynofloras show an increase in S/D ratio throughout

the *Cucullaea* II Member (DJ 804.14-16, 807.3-16) as the floras become more terrestrially-influenced.

The uppermost member of the La Meseta Formation, the Submeseta Member, also shows increasing S/D ratio values across its boundary with the underlying *Cucullaea* II Member. It then displays a drop in S/D ratio, indicating an increase in marine influence, however, the majority of the thickness of this member is unsampled. It is therefore impossible for any rigorous analysis to be undertaken on this unit.

Interpretation of the S/D ratio data for the La Meseta Formation indicates that a shallow-marine setting is most probable, as has been suggested by previous reconstructions (Wrenn & Hart, 1988, Porebski, 2000). The overall trend through the formation is for a move from marine-dominated assemblages to more terrestrially-dominated assemblages. This is indicative of a shift in depositional environment from a more distal setting to one much closer to the shoreline, e.g. a shallowing upwards, or deposition further up inside the La Meseta estuary. The shallowing upwards trend is seen primarily in the upper La Meseta Formation (*Cucullaea* I – Submeseta members) with little evidence of it in the lower La Meseta Formation (Valle de las Focas – Campamento members). This is in line with the assertion of Wrenn & Hart (1988 p. 342) that the 'marine environment became progressively shallower upsection.' The distinction between the lower three members and the upper three members of the La Meseta Formation is also in line with the suggestion of Porebski (2000) that this boundary represents a major eustatic sea level low at 49.5Ma. A hiatus of this magnitude could allow significant changes in the depositional environment to develop prior to the resumption of sedimentation. The continued shallowing upward trend seen in the upper members of the La Meseta Formation

following the postulated 49.5 Ma low, is more likely to be related to localised changes in relative sea level within the palaeovalley, than be related to globally correlated eustatic changes.

The overall trend through the La Meseta Formation, as mentioned above, is for a shift from more marine-dominated floras to more terrestrial floras, probably as a consequence of infilling of the incised valley estuary system within which these sediments were being deposited. Within this are smaller-scale changes. Fluctuations within the individual members can be attributed to the large variation in environments and energy settings within an estuary, these were described in the La Meseta estuary by Porebski (1995), who described the separate facies and their depositional environments. Differences between the individual members, e.g. across unconformable boundaries, reflect relative sea level change, caused by eustatic sea level changes, localised tectonism, sedimentary infill, or a combination of these.

The P/G ratio curve (Fig. 4.5B) shows an overall similar trend to that of the S/D ratio analysed above. The upper part of the La Meseta Formation shows an overall increase in the productivity index. In the lower part of the formation, although there are considerable fluctuations there does not appear to be an overall trend for the assemblage proxy to increase or decline in productivity.

The basal Valle de las Focas Member shows very high levels of peridinioid dinoflagellate cysts, this is possibly as a consequence of their presence in the uppermost section of the Sobral Formation that forms the lower part of this section, e.g. being reworked into the younger sediments. A rapid decline in P/G ratio is apparent once sedimentation within the Valle de las Focas Member is established, prior to the sampling gap (DJ 801.2-5). The uppermost Valle de las Focas (DJ 801.8)

exhibits a marked increase in the P/G ratio, indicating a rise in productivity during the unrepresented section.

Productivity declined across the boundary with the Acantilados Member and productivity remains generally steady throughout the member until a dramatic increase in P/G ratio in the uppermost Acantilados Member (DJ 802.13-1, 803.1-5). A secondary feature in the P/G curve throughout the Acantilados Member is a thickness of elevated productivity (DJ 802.30-24), though it still remains relatively unproductive compared to other parts of the curve. However, this feature is also coincident with the broad-based excursion described above for the S/D curve (Fig. 4.5A). This suggests a link between higher levels of terrestrial influence and higher productivity, such as a greater nutrient input from the land.

In the Campamento Member, the P/G ratio shows a similar character to that seen in the underlying members, e.g. productivity drops sharply across the boundary. It then rebounds immediately to higher productivity levels before reaching the sampling gap. Again, this makes it very difficult for a reliable analysis or interpretation to be done on this part of the section.

In the upper part of the La Meseta Formation, the overall trend is one of an increase in the P/G ratio through *Cucullaea* I and *Cucullaea* II. Both members have very low initial P/G ratio values and both show a subsequent increase in productivity throughout their thickness, although considerable variation is seen within this overall increase, especially in the *Cucullaea* II Member. The Submeseta Member shows a sharp decline in P/G ratio in the small thickness represented. No trend can be recognised and no interpretations can safely be made from such an unrepresentative curve.

As mentioned above, there is a similarity in the overall trends seen in both the P/G and S/D (Fig. 4.5A) data. The incidence of assemblages displaying both elevated S/D ratios and P/G ratios, e.g. DJ 802.29-23 & DJ 904.1-807.16 suggests that the shallower waters in the La Meseta estuary have greater nutrient levels than the deeper waters. This may be due to the embayment of shallower waters within the estuary, allowing the concentration of peridinioid dinoflagellate cysts in nutrient-rich waters. The deeper waters may have been open causing the diffusion of the nutrient-load and creating conditions less favourable to the peridinioid dinoflagellate cyst community. Interpretation of the productivity index data also indicates that each member displays the trend of increasing the proportion of peridinioid dinoflagellate cysts towards the top of each section, indicating an increase in water productivity. Each member also has a low P/G ratio in the lowermost sediments indicating low productivity conditions. This links to the above, in that each member shows an increase in productivity as it shallows upwards, moving closer to the shoreline and further up the estuary valley during each period of fill.

Analysis of the species richness plot (Fig. 4.5C) shows that the overall trend through the La Meseta Formation is of a rise in species richness from the base of the formation, to a peak in the lower Acantilados Member (DJ 802.24-3). Species richness then declines through to the top of the La Meseta Formation. There is, however, variation seen within each member. Among the members with a complete record of species richness, the Acantilados and *Cucullaea* II members both show an initial low species richness, increasing to higher levels in the lower or middle part of the sediment thickness. This is followed by a decline to the top of the member. The *Cucullaea* I Member also displays this overall trend, but has an initial peak of high species richness. This could be due to the effect of condensation, which is a feature of

transgressive lags (Kidwell, 1989), such as that from which this sample was taken (Fig. 4.5).

The species richness data can be interpreted to show a record of the incision, deepening and sediment infilling represented by each individual member throughout the La Meseta Formation. The species richness (see above) can be used as an indicator of water depth and it is possible to see within the displayed data evidence of an increase and then a decrease in relative sea level in each member. When compared to the curve generated in Fig. 4.5A, the proxy to estimate distance to the shoreline, the overall trends are the same. The most obvious feature within the species richness data is a section of elevated species richness (DJ 801.8-802.23) in the Acantilados member. This coincides with the excursion seen in both Fig. 4.5A-B and suggests that the highest species richness of all is not related to increasing relative sea level and distance to the shoreline, but to nearer shore, high-productivity conditions.

The absolute abundance curve (Fig. 4.5D) displays an overall trend similar to that seen in the species richness curve (Fig. 4.5C), e.g. an increase in abundance into the lower Acantilados Member followed by a decline to the top of the formation with secondary features apparent within the curve. The most significant feature within the curve is the high levels of marine palynomorph absolute abundance seen in the lower Acantilados Member (DJ 802.33-29). This is followed by a sudden decline in abundance (DJ 802.28-24) in the upper part of the lower Acantilados Member. Following this decline, abundance levels recover briefly to higher levels (DJ 802.23) before declining through to the top of the member. Other features within the abundance curve include a peak of abundance in the lowermost *Cucullaea* I Member (DJ 804.1) followed by lower, but still elevated abundance levels (DJ 804.4-8). Following abundance highs in the lower *Cucullaea* I Member abundance declines

sharply and remains at low levels through the rest of the La Meseta Formation (*Cucullaea* II and Submeseta members).

Interpretation of the absolute abundance results suggest a correlation between the high levels of absolute abundance seen in DJ 902.33-29 with high levels of marine influence (Fig. 4.5A) and low levels of productivity (Fig. 4.5B). These proxy results are indicative of a setting with higher relative sea levels and low productivity. The lower levels of abundance within the Acantilados Member, conversely, correlate with the higher levels of terrestrial influence, eutrophic conditions and high species richness described above. It is possible that unusual conditions existed at this time, as usually these parameters are linked to high abundances. The decline in absolute abundance in the upper part of the Acantilados Member is compatible with the increase in terrestrial influence and a shift towards the shoreline. The lowermost *Cucullaea* I Member is characterised by an abundance peak that correlates with a similar high species richness (Fig. 4.5C). The Campamento-*Cucullaea* I boundary is described as a sequence boundary (Porebski, 2000, see above) and the lowermost sediments are described as a transgressive lag (Porebski, 1995). In this case, the high levels of both abundance and diversity could be as a consequence of stratigraphic condensation (Kidwell, 1989). The levels of abundance and species richness through the rest of the Cucullaea I Member do correlate and they show a decline through to the top of the section. This also corresponds to the increase in the S/D ratio (Fig. 4.5A). The low absolute abundance data shown from the uppermost *Cucullaea* I to the top of the Submeseta Member is unexpected given the sedimentological evidence (Porebski, 1995). It has been reported by Porebski (2000) that there is a basinward facies shift following the boundary between the *Cucullaea* II and Submeseta members. Following this, facies become more fluvially-dominated. This may have

some influence on the decline in abundance, however, the decline is seen before this boundary so some other mechanism must be sought. Cantrill (pers. comm.) describes the facies variation as remaining consistent throughout the La Meseta Formation, so it may be more likely that the decline in abundance has a post-depositional cause and may instead by due to exposure and oxidation following the uplift and exposure of the sediments containing the floras.

Analysis of the stratigraphic distribution of individual species and taxa (Fig. 4.6) reveals details of the changes both through and within the La Meseta Formation.

The relative abundance of individual species and taxa (Fig. 4.6) shows trends similar to those already seen (Fig 4.6). These are primarily due to the relative abundance distribution of *Enneadocysta partridgei* and *Enigmadinium cylindfloriferum*, which dominate the marine assemblages of the La Meseta Formation. Other taxa that are dominant in the formation include *Octodinium askiniae, Vozzhennikovia apertura* and *Spinidinium* spp. (including *S. macmurdoense* and *S. essoi*). The most obvious feature is the changing dominance between *E. partridgei* and *E. cylindfloriferum* in the Acantilados Member. Other features apparent are the disappearance of *E. cylindfloriferum* in the upper Acantilados Member, the continued domination of the assemblage by *E. ursulae* and the increasing abundance of *V. apertura* in the upper La Meseta Formation.

Enneadocysta patridgei (as *Areosphaeridium* cf. *diktyoplokus*) was reported by Cocozza & Clarke (1992) to be usually associated with open marine conditions. However, Askin *et al.* (1991), has suggested that the elaborate processes characteristic of this species may be a functional adaptation to the abrupt changes in salinity seen where fresh water and sea water meet, e.g. hypopycnal conditions. The highest

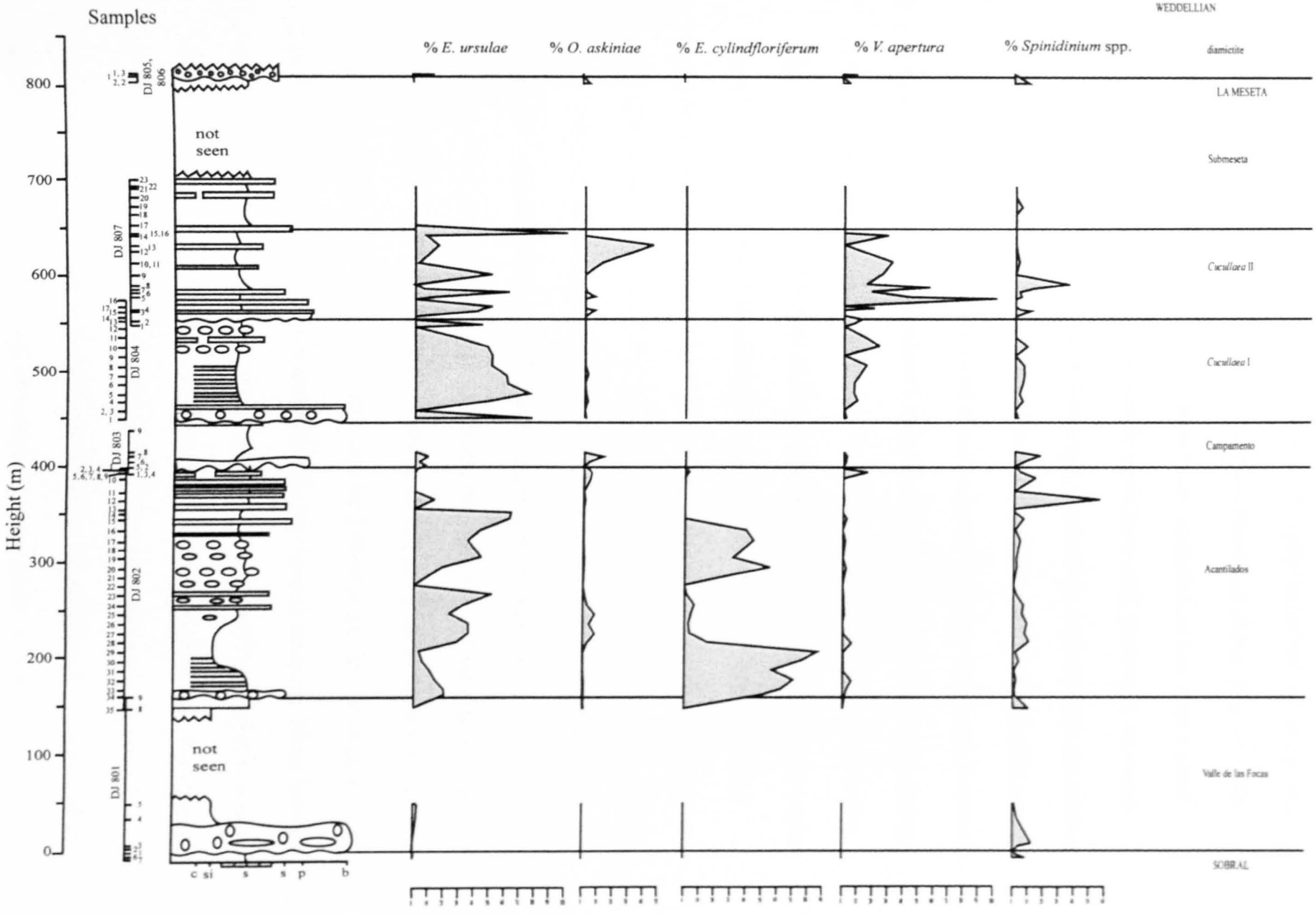


Fig. 4.6 Diagram of dominant marine palynomorphs species of the La Meseta Formation (as re. %)

abundances of *E. partridgei* are seen in DJ 802.28-23 and again in DJ 802.20-14. These abundances correspond to the intervals of higher terrestrial influence, higher productivity and higher species richness already described above (Fig. 4.6). The environments recorded in these sediments would appear to be less stressed and possibly lower energy. The increase in the abundance of *E. partridgei*, so closely connected to lower relative sea level and high productivity indicates that the linking of this species by Askin *et al.* (1991) to conditions that would be found within an estuary is correct. The distribution of *E. partridgei* in the upper La Meseta Formation is initially similar: in the *Cucullaea* I Member its presence is again linked to strong marine influence, high productivity, and high species richness. However, from the uppermost *Cucullaea* I Member and upwards the distribution decline becomes more erratic. This can be linked to a further increase in terrestrial influence, possibly creating conditions unsuitable for *E. partridgei*. Species richness also declines at this point (Fig. 4.6).

No environmental interpretations for *E. cylindfloriferum* have been made, although Cocozza & Clarke (1992) have suggested a link to stressed palaeoenvironments. The highest relative abundances of *E. cylindfloriferum* are seen in DJ 802.34-29 with a second, lower relative abundance peak in DJ 802.20-16. The highest abundances are linked to a strong marine influence and low productivity as well as high absolute abundances and relatively low species richness. These indicate stressed conditions, in agreement with Cocozza & Clarke (1992), possibly in a wave-dominated, outer part of the estuary, given the strong marine influence. The absence of *E. cylindfloriferum* in the upper part of the La Meseta Formation suggests that either the stressed, marine-influenced conditions no longer existed in this part of the palaeovalley, e.g. deposition was occuring more proximally, or that *E.*

cylindfloriferum had become extinct, or both. Wrenn and Hart (1988) have dated the species extinction to late Early Eocene and its disappearance at the top of the lower La Meseta Formation would match the placement of the 49.5 Ma sea-level lowstand at the top of the Campamento Member (Porebski, 2000).

The other selected taxa are less dominant then the two described above, however, *V. apertura* does become an important part of the flora in the upper part of the La Meseta Formation. The distinction between its distribution in the lower and upper sections of the formation suggests that there is a difference in environment between the two sections and that this could well be as a consequence of the sedimentary hiatus proposed both by Wrenn & Hart (1988) and Porebski (2000). The occurrence of *O. askiniae* appears to be erratic and random and there is no discernible pattern to its distribution. Wrenn & Hart (1988) suggest that it may be reworked and that seems likely. The *Spinidinium* complex of species present also lack a consistent pattern of abundance, although there may be some correspondence between their presence and the high relative abundance of *E. partridgei* in the Acantilados (DJ 802.28-23 & 802.20-14) and *Cucullaea* I (DJ 804.4-11) members.

4.1.2 - Correspondence Analysis

Correspondence Analysis was run on a matrix of the 37 most abundant species in the López de Bertodano and Sobral formations (formations lumped together for this analysis), and of the 31 most abundance species in the La Meseta Formation. Other selections of species were made to check for consistency. The scatterplots of loadings of samples (Fig. 4.7A) and species (Fig. 4.7B) on the first two correspondence axes are shown which allows comparison of the clustering of sample loadings and the species responsible for these results. Additionally, eigenvector

loadings of samples are plotted against depth (Fig. 4.8) to permit analysis of stratigraphic variation and easier comparison with previous analyses.

4.1.2.1 - López de Bertodano Formation

Correspondence Analysis results for the dinoflagellate cyst flora are dominated by a clear floral turnover in the upper part of the López de Bertodano Formation (Fig. 4.7A). The majority of samples from the López de Bertodano Formation are separated from the overlying Sobral Formation by the first Correspondence Analysis (CA), which accounts for 14.81% of the assemblage. Virtually all species representing the López de Bertodano Formation contribute moderately-strongly to positive loadings on the first CA (exceptions are DJ 904.43-46 & DJ 904.71-2). Sample loadings from the same species on the second CA, which accounts for a further 10.88% of the assemblage, show a wider distribution, with both positive and negative loadings.

Loadings on the first CA plotted against depth show a clear trend through the López de Bertodano Formation (Fig. 4.8). The most obvious feature of the curve is a steady increase in the eigenvector score loadings through the majority of the sampled thickness (DJ 904.44-68). *Trithyrodinium evitti* contributes the strongest positive loading on the first CA. Other distinct features include a small cluster of the basal four samples (DJ 904.39-42) before a marked decline (DJ 904.43-44) to the beginning of the major positive excursion trend described above.

Further marked decline in score loadings is seen at the top of the positive excursion, in the uppermost part of the López de Bertodano Formation (DJ 904. 904.71-2). This can be interpreted as a low diversity basal assemblage in the earliest Paleocene seen in the lower four samples, dominated by the species *Subtilisphaera*

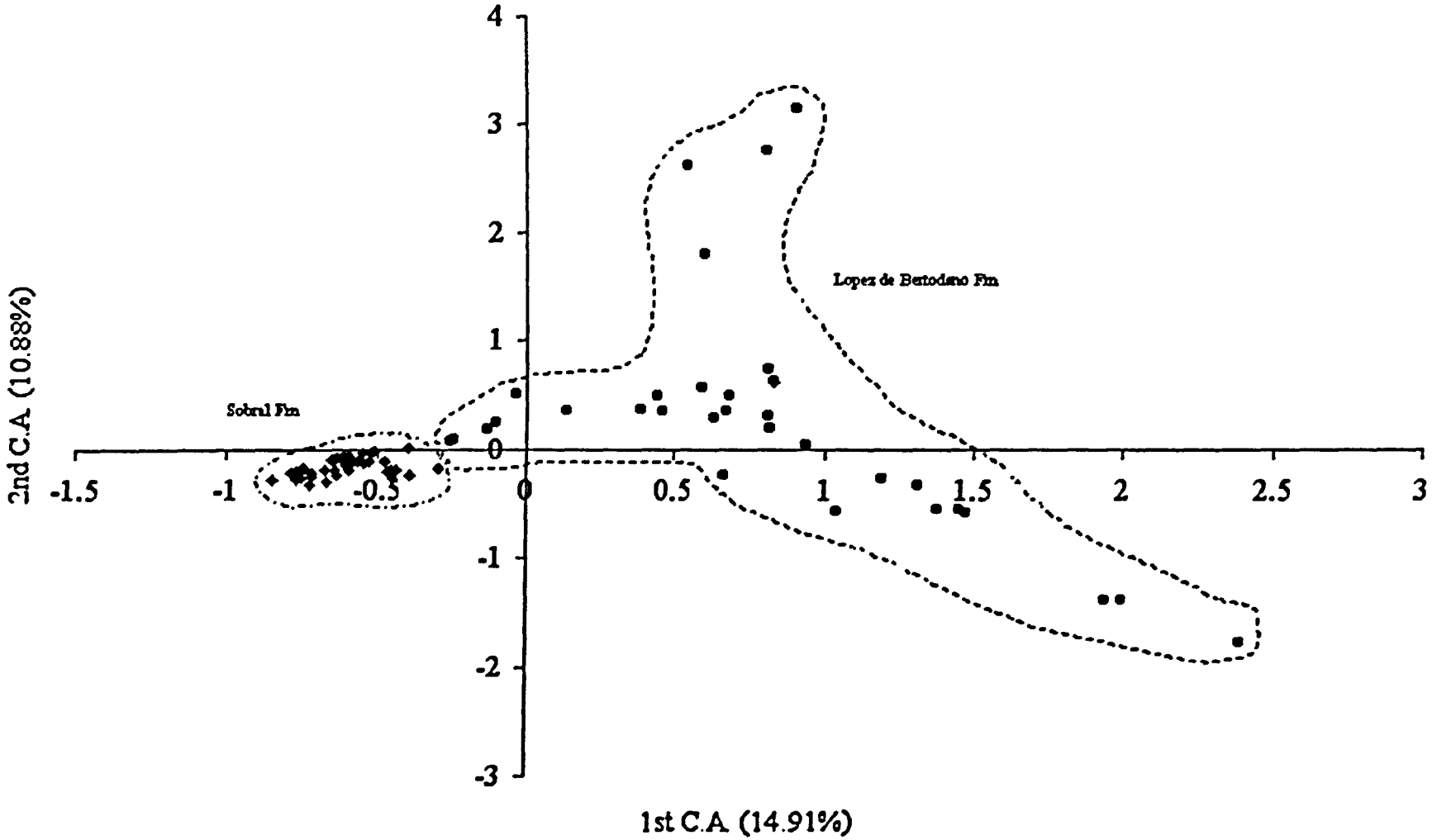


Fig. 4.7A Loadings of samples of the Lopez de Bertodano and Sobral formations

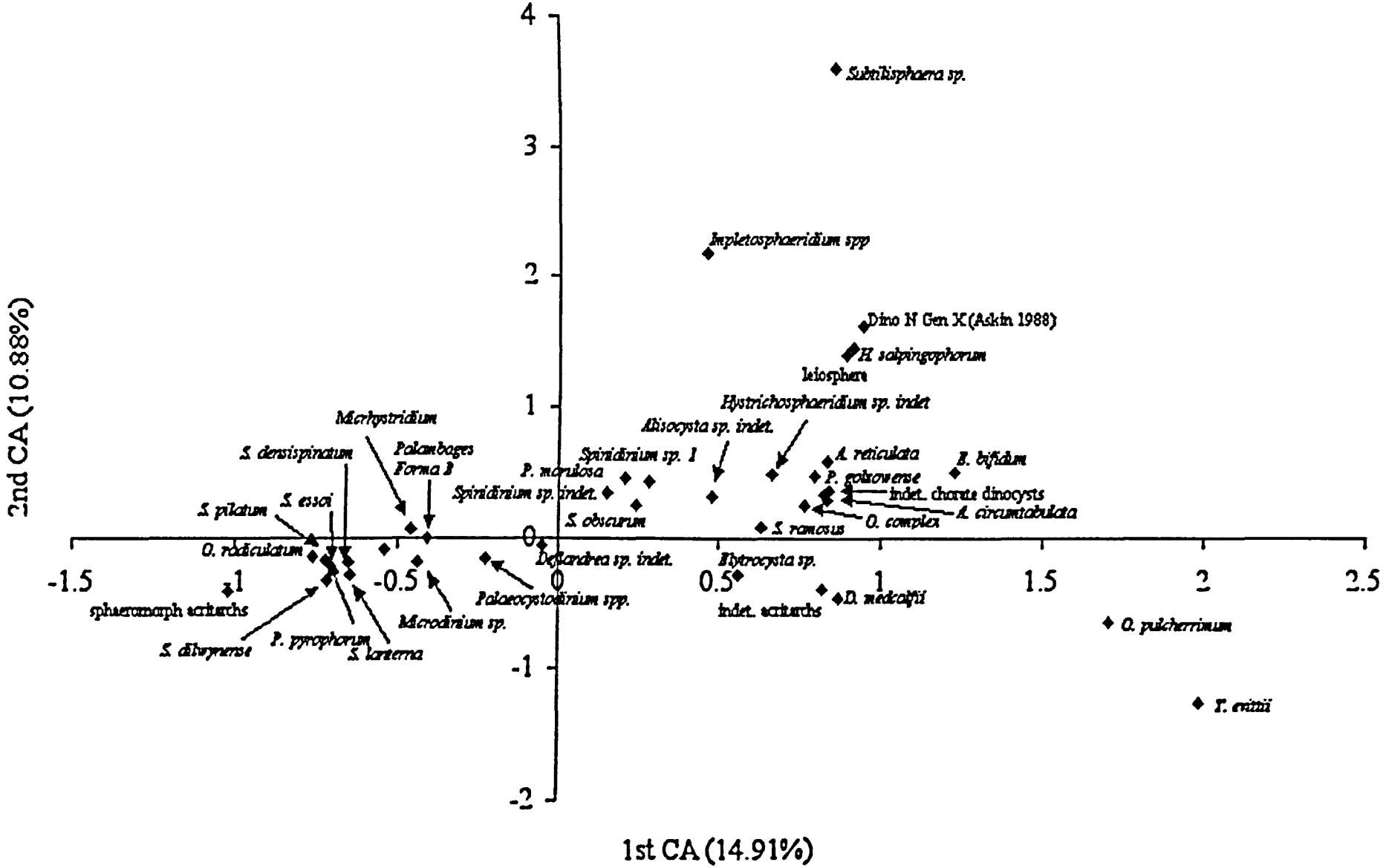


Fig. 4.7B Loadings of Species of the Lopez de Bertodano and Sobral formations

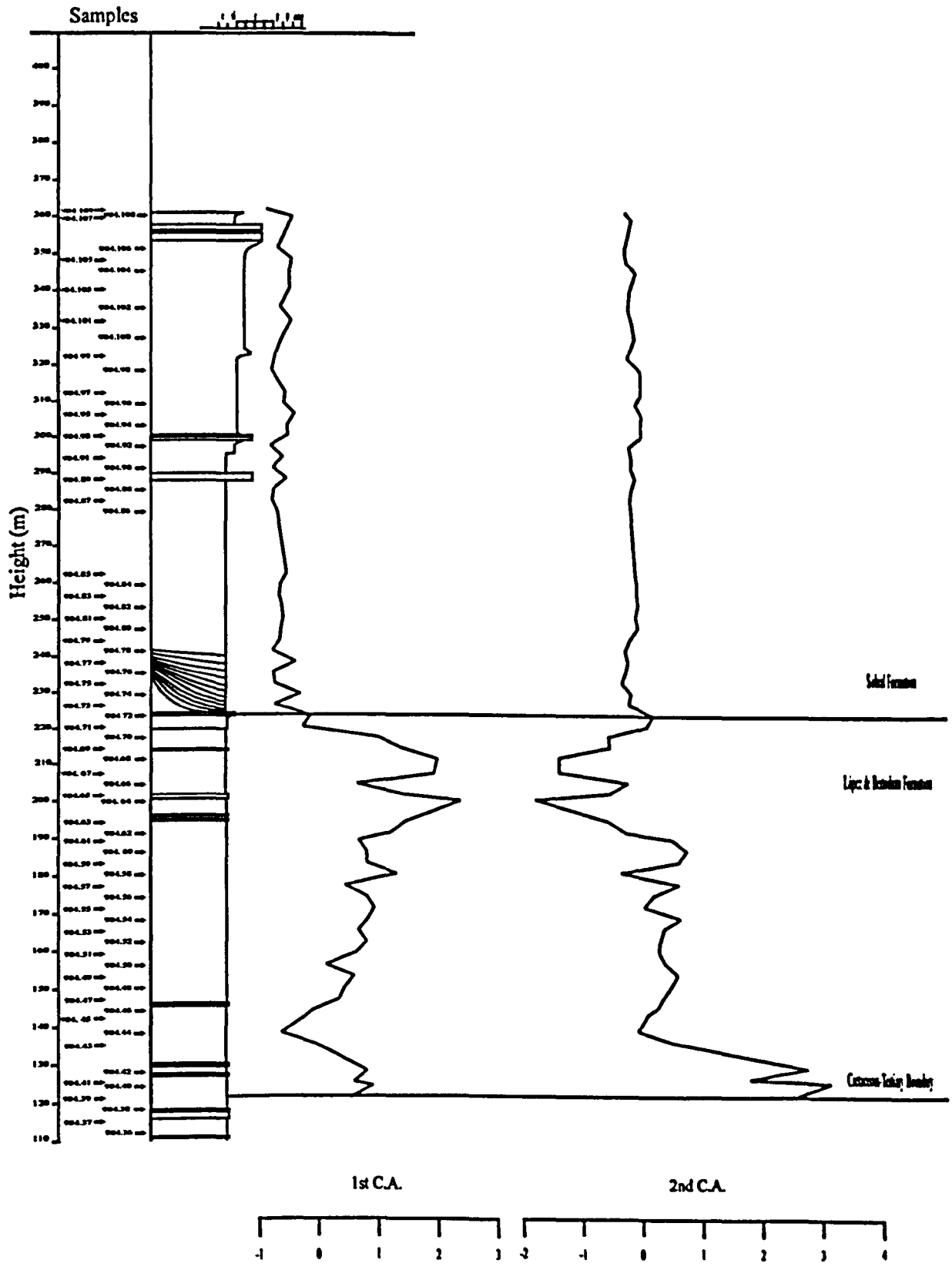


Fig. 4.8 Loadings of the first and second Correspondence Axes plotted against depth

spp., *Impletosphaeridium* spp., Dino N Gen X (Askin, 1988), *Hystrichosphaeridium salpingophorum* and leiospheres. There is then a small transition flora to the base of the long-term increasing trend which is characterised by a disappearance of the above species and the brief appearance of *P. pyrophorum* and *Micrhystridium* spp., which indicates changing conditions. The presence of the heterotrophic *P. pyrophorum* indicates an increase in productivity and possible eutrophic conditions. The positive eigenvector loading scores which characterise the largest part of the samples from the López de Bertodano Formation are caused by the presence of *Alisocysta reticulata*, *Alisocysta circumtabulata*, *Deflandrea medcalfii*, *Exosphaeridium bifidum*, *Oligosphaeridium complex* and *Oligosphaeridium pulcherrimum*, and the highest loading scores are contributed by *Trithyrodinium evitti* (see above) (Fig. 4.7B). *T. evittii*, like *P. pyrophorum*, is a heterotrophic species. Its dominance of this assemblage would require high levels of nutrients to be available, probably as a consequence of terrestrial input or offshore upwelling. The uppermost samples showing a decline in score loadings as a consequence of the decline in the dominance of the above species. The changing assemblage composition becomes closer to that of the overlying Sobral Formation, suggesting that the changes in the depositional were already apparent in the López de Bertodano Formation. Heterotrophic species, e.g. *P. pyrophorum*, contribute to score loadings so it is apparent that productivity levels remained high, whatever other changes must have occurred.

Loadings on the second CA (Fig. 4.8) plotted against depth show a very similar pattern, although there is a higher degree of variation in the loading scores. The same trends are visible, with the same samples grouped together.

4.1.2.2 - Sobral Formation

Correspondence Analysis results for the dinoflagellate cyst flora are dominated by a striking homogeneity throughout the samples present. They show a distinct separation from the majority of the underlying López de Bertodano Formation, although there is some correspondence with the uppermost samples of this formation. All sample score loadings have a weak negative loading with little variation on the first CA (Fig. 4.7A), which accounts for 14.81% of the assemblage. On the second CA, which accounts for 10.88% of the assemblage, there is less variation and all score loadings still have a negative contribution.

Loadings on the first CA plotted against depth (Fig. 4.8) show no overall trends, although there are two areas showing more pronounced variation against the background. The basal samples of the Sobral Formation (DJ 904.73-78) show variation across the stratigraphic interval deposited as the Sobral channel eroded into the top of the López de Bertodano Formation sediments. This would have been a stressed environment. Score loadings are contributed to by *Palambages* Forma B, a marine algae, and *Palaeocystodinium* spp, a heterotrophic genus requiring significant nutrient levels. These are not incompatible with stressed environments. The second area of variation occurs higher in the section (DJ 904.79-93). These samples correspond to an interval containing a triple glauconite band and increasing volcanic detritus and wood (D. Cantrill, pers. comm.). Glauconite is deposited in marine environments with low sedimentation rates, consequently stratigraphic horizons can become condensed and this can affect the planktonic fossils they contain. The lack of sedimentation suggests a change in the local environment.

Loadings on the second CA plotted against depth (Fig. 4.8) again show no overall trends, although the pattern of variation differs slightly. The basal variation in DJ 904.73-78 is still apparent, but subsequent samples (DJ 904.79-93) show a marked lack of variation. This interval of little variation comes to an end at the triple glauconite bed mentioned above, at which point sample score loadings again begin to show some variation. The change at the glauconite level indicates a difference between the two units. The condensed interval may indicate a sequence break, e.g. a maximum flooding zone, or a more localised change with sediment bypassing the immediate area.

The overall lack of any significant change in either CA suggests a generally stable environment with few changes in the dinoflagellate cyst population.

4.1.2.3 - Cross Valley Formation

Samples from the Cross Valley succession did not contain a dinoflagellate cyst flora, therefore no further analysis could be made.

4.1.2.4 - La Meseta Formation

Correspondence Analysis results for the dinoflagellate cyst flora show considerable differences in loadings of samples between the different members of the La Meseta Formation. . The majority of samples show little variation on the first CA (Fig. 9A), which accounts for 12.14% of the assemblage. The species in these samples have a weak negative or positive loading, but there are also several samples which contribute very strong positive loadings. Sample loadings on the second CA (Fig. 9A), which accounts for 9.3% of the assemblage, show a higher amount of variation among the samples. The species these samples contain contribute a moderately positive to

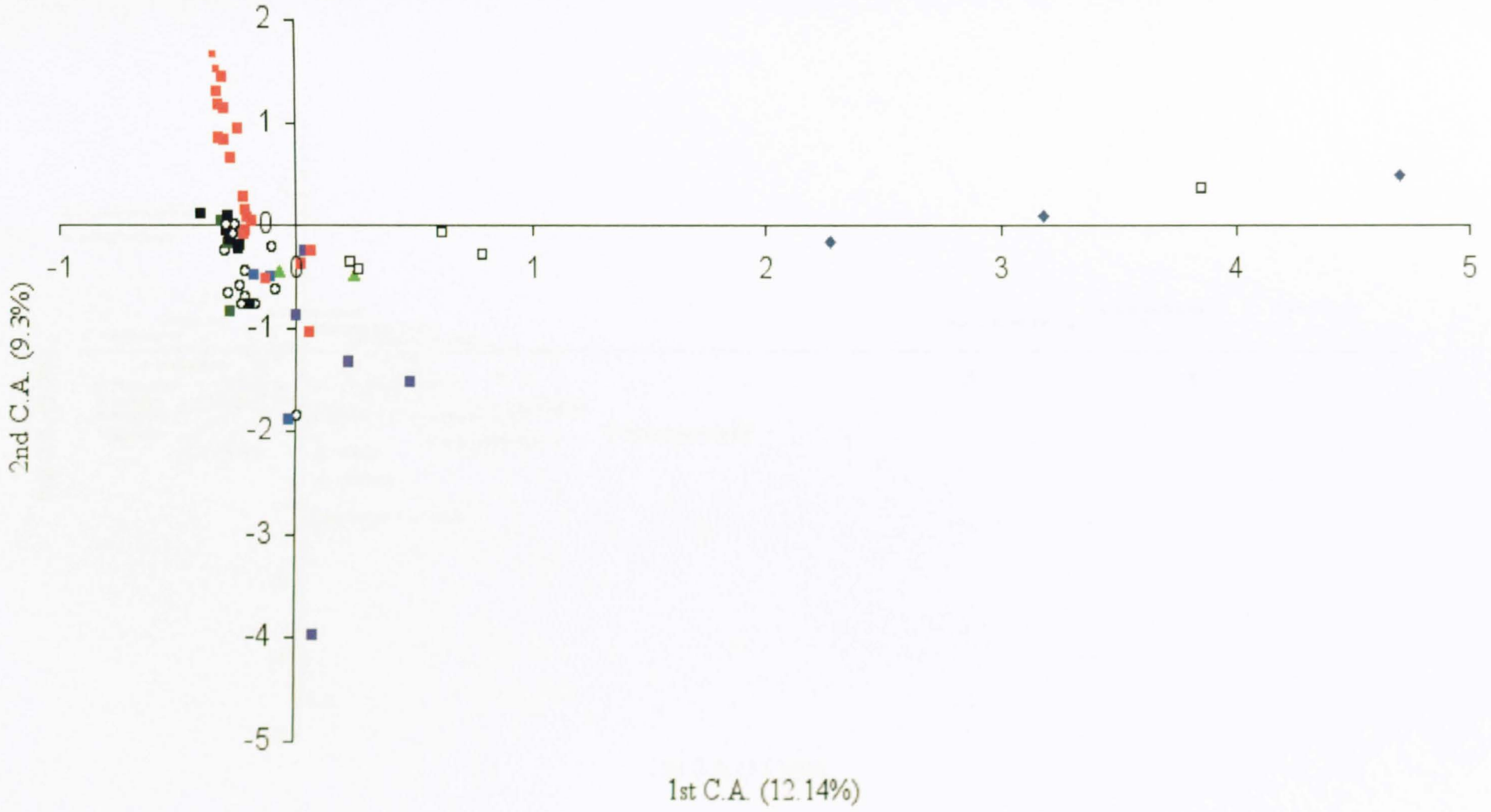


Fig. 4.9A Plots of loadings of samples of the La Meseta Formation

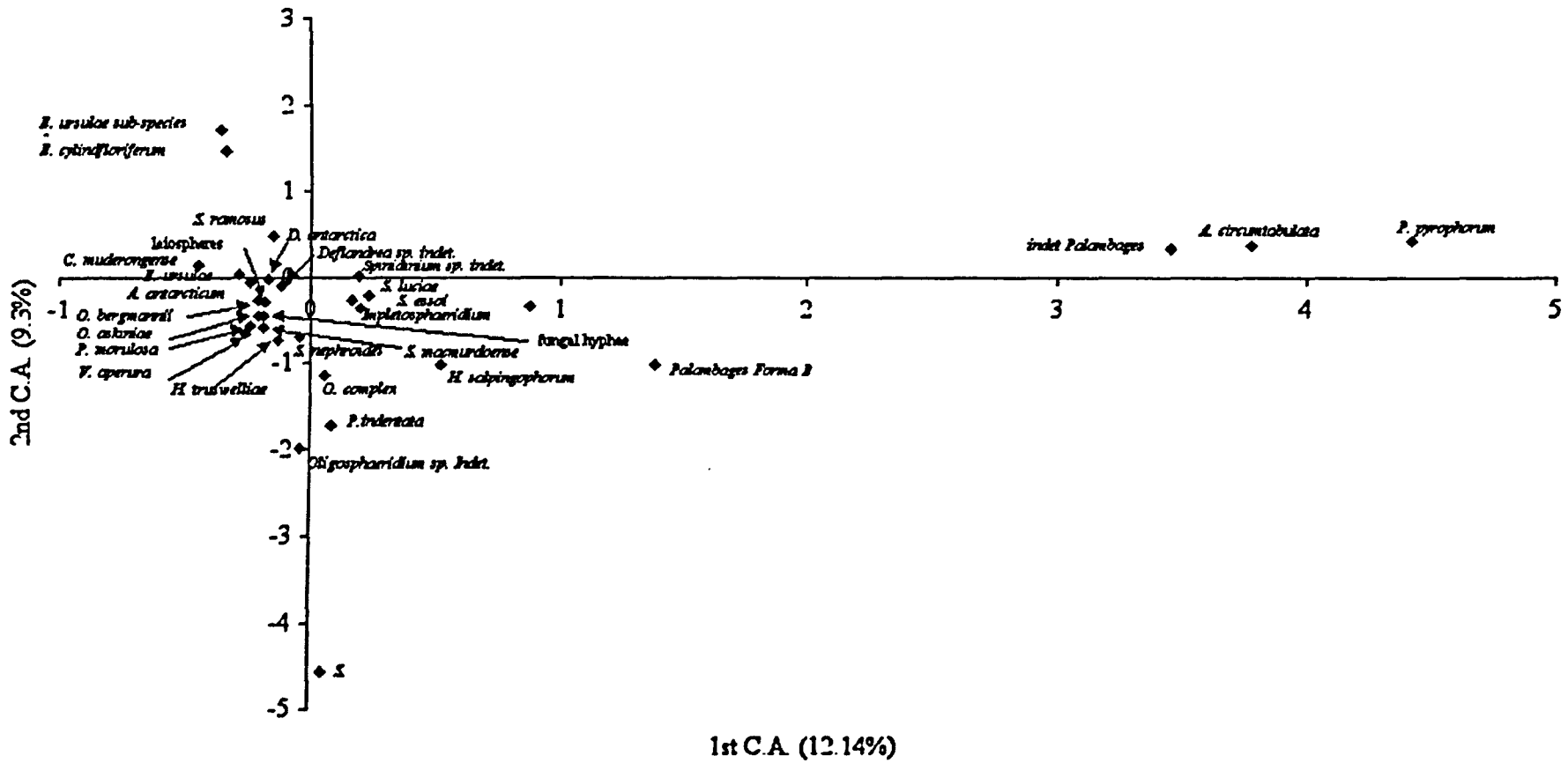


Fig. 4.9B Plots of loadings of species of the La Meseta Formation

weakly negative loading, with one outlier with a strongly negative loading (Fig. 4.9A).

Loadings on the first CA plotted against depth (Fig. 4.10) show little variation through the majority of the La Meseta Formation, most samples display a weakly negative loading. The obvious feature of the loading curve is the strong positive excursion at the base of the formation (DJ 801.1-2 & 801.6-7). This is seen at the contact between the uppermost Sobral Formation (DJ 801.1 & 801.6-7) and the basal La Meseta Formation (DJ 801.2), in the lower part of the Valle de las Focas Member. *P. pyrophorum* contributes the strongest positive loadings (Fig. 4.9B), along with *A. circumtabulata* and indeterminate *Palambages* sp. A secondary feature present is a small positive excursion in the Submeseta Member (DJ 807.19-22) caused by the loadings of the species *Semispinula ancoriferum* and *Oligosphaeridium* spp.

The basal positive excursion can be interpreted as reflecting the uppermost Sobral Formation, with consequent floral reworking as a result of being eroded and crosscut by the younger La Meseta Formation. The beds of the Sobral Formation are significantly older than those in the overlying sediments and consequently have a different flora, as is seen in the CA results (Fig. 4.9). The elevated positive loadings for the basal La Meseta Formation (Fig. 4.9-1), the Valle de las Focas Member, can be interpreted by the continued reworking of these older species into Eocene deposits due to the levels of erosion caused by the incising of the La Meseta valley. Higher in the Valle de las Focas Member, sample loadings are reduced as the assemblage becomes more dominated by *Spinidinium* spp., and *Impletosphaeridium* spp. The presence of the positive excursion in the Submeseta Member of *S. ancoriferum* and *Oligosphaeridium* spp., both reworked from the Late Cretaceous, can be interpreted

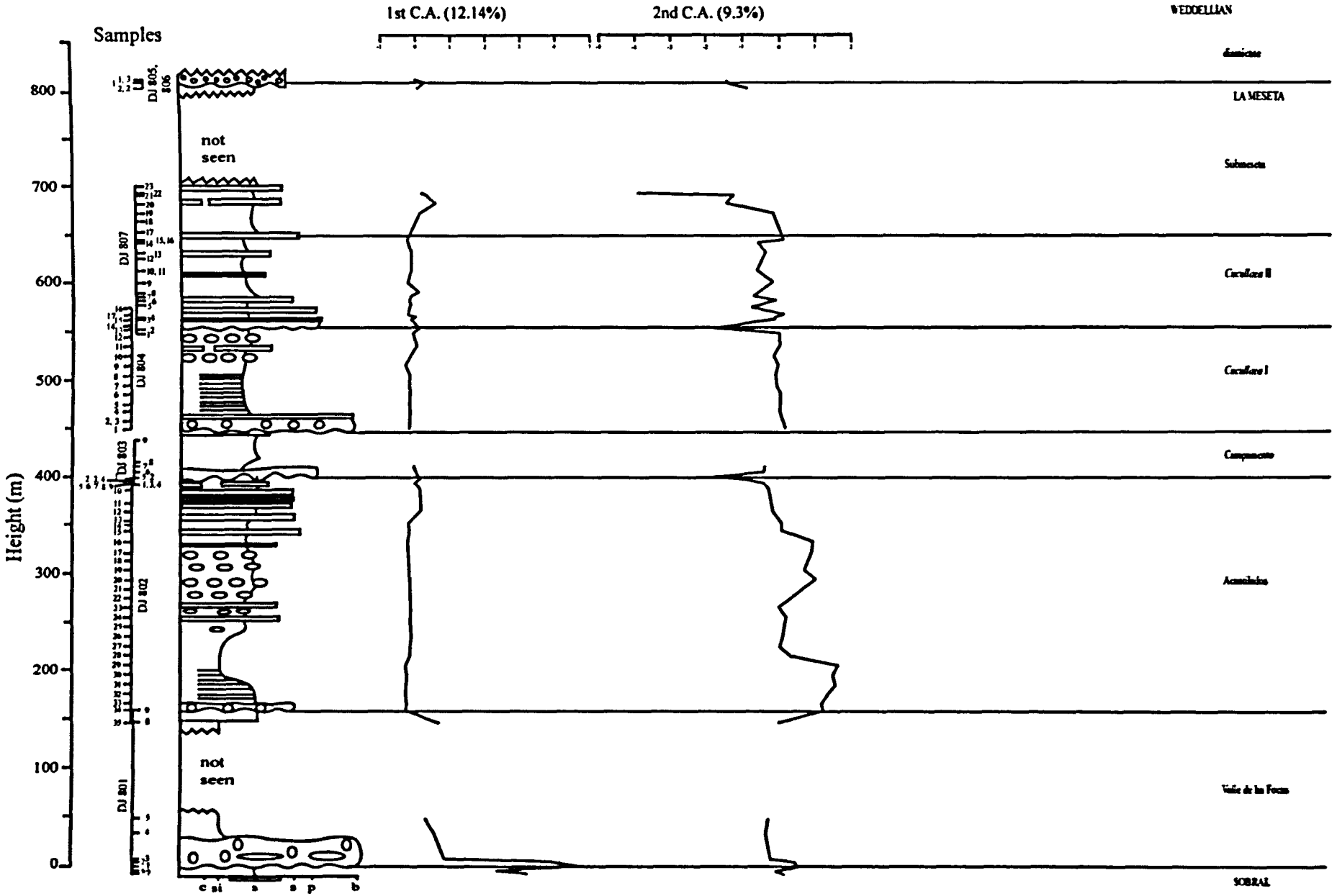


Fig. 4.10. Score loadings of the first and second Correspondence Axes plotted against depth

as being due to a further period of erosion. This could the consequence of either subaerial exposure, the tidal scouring of older sediments or the erosional base of a channel. The rest of the La Meseta Formation, e.g. the Acantilados, Campamento, *Cucullaea* I and II members, show very little variation or strong contribution in their sample loadings. The taxa contributing to these samples include *Arachnodinium antarcticum, Deflandrea antarctica, Enneadocysta partridgei, Octodinium askiniae, Palambages morulosa, Spinidinium macmurdoense* and *Vozzhennikovia apertura.*

Loadings on the second CA plotted against depth (Fig. 4.10) produce a different pattern. There are strong negative and positive excursions apparent through the section, although there does not appear to be an overall trend. The most obvious feature of the curve are two broad positive excursions in the Acantilados Member. The first of these excursions (DJ 801.9/802.34-802.29) begins at the base of the Acantilados Member and extends for 50m. The second excursion occurs in the mid-upper part of the member (DJ 802.21-16) and also extends for around 50m. The species loadings causing these elevated scores, in both cases, are *Enneadocysta partridgei* sub-species 1 and *Enigmadinium cylindfloriferum* (Fig. 4.9B). Other major features of the curve are the three negative excursions. The first of these occurs in the basal Campamento Member, immediately above the contact with the underlying Acantilados Member (DJ 803.5/803,2/802.1). This excursion appears as a single peak and the sample loadings are caused by the presence of *Oligosphaeridium* spp. and *Paralecaniella indentata* in the assemblage. The second negative excursion is very similar, occuring in the basal *Cucullaea* II Member, immediately above the contact with the underlying *Cucullaea* I Member (DJ 807.2/804.13). The sample loadings appear to be caused by the presence of the same species, *Oligosphaeridium* spp. and *P. indentata.* The third negative excursion occurs in the Submeseta Member and is

equivalent to the small positive excursion of the first CA (DJ 807.19-22) described above.

The palaeoenvironmental implications of the presence of *E.partridgei* subsp. A are unclear, while E. *cylindfloriferum*, as seen above (4.2.1.4) is likely to represent stressed conditions and the clear separation of this taxon from the rest of the assemblage suggests that it has very different environmental preferences. The interpretation of the negative excursions occurring in the basal Campamento and Cucullaea II members is more straightforward. Both these assemblages are part of the basal deposit of a transgressive lag. Transgressive lags are recognised to contain a higher proportion of reworked material (Kidwell, 1989) as they are the part of the first significant marine flooding surface to be deposited across the shelf during a transgression (Emery & Myers, 1996). The negative excursion in the Submeseta Member is not linked to a transgressive lag, however it is linked to a sandstone body that could represent a smaller channel deposit. The basal part of such a channel presents a similar pattern to a flooding surface, as it is an erosional event. As in the first CA (Fig. 4.10), the majority of the La Meseta curve does not express any significant sample loading. The species responsible for this loading are the same as listed above, e.g. *Arachnodinium antarcticum*, *Deflandrea antarctica*, *Enneadocysta partridgei*, *Octodinium askiniae*, *Palambages morulosa*, *Spinidinium macmurdoense* and *Vozzhennikovia apertura*. It is possible that these species represent quieter environments of deposition that were disrupted by the higher energy environments of the basal sections of each member.

4.2 - CONCLUSIONS

The aim of this study was to study high resolution sea level changes and environmental fluctuations as reflected in changes in the marine palynofloras recovered from the shallow marine sediments of the Palcogene sections of the Antarctic Peninsula.

The marine palynofloral record in the Paleocene shows the palaeoenvironment of the López de Bertodano Formation to have been a shallow marine setting, with evidence for a regressive trend through much of the measured section There is also evidence for a corresponding transgression in the uppermost interval of the formation. The waters were moderately to highly productive, with an apparently terrestrial source for the nutrients.

The dinoflagellate cyst assemblage living within this environment was predominantly peridinialean, although the lowermost flora is more mixed. The samples contain an assemblage dominated first by *P. pyrophorum* and *Spinidinium* spp., and then by *T. evittii.* This turnover represents an influx of warm water from lower latitudes (Brinkhuis *et al.*,1998), characterised by the low latitude species *T. evittii.* The dominance of *T. evittii* is not long-lived, and the assemblage reverts to its previous composition, reflecting a change back to cooler water temperatures.

The Sobral Formation is separated from the underlying López de Bertodano Formation by a localised hiatus up to 100m and of apparently little duration in time. Shallow marine conditions are suggested; the lower part of the section is represented by the top of a transgressive cycle that began in the upper López de Bertodano Formation, while the rest of the measured section reflects a regression that continues through the top of the measured section. The waters were consistently highly

productive throughout the section present, indicating eutrophic conditions with little variation in nutrient availability.

The dinoflagellate cyst assemblage living within this environment was predominantly peridinialean, *P. pyrophorum* was the most abundant individual dinoflagellate cyst species present with *Spinidinium* spp. another important taxa. The shallow water acritarch genus *Micrhystridium* spp. was also a major part of the flora during the transgressive sub-cycle in the lower part of the formation. Both species richness and abundance declined through the measured interval of the Sobral Formation, again reflecting the regressive trend discussed above.

The Cross Valley Formation contains negligible marine floras.

The Eocene La Meseta Formation lies unconformably over the Paleocene Sobral Formation into which the younger sediments are deposited within a fault-bounded valley. The samples from the La Meseta Formation show the palaeoenvironment to have been an overall shallow marine setting, with a further shallowing up trend present in the upper three members (*Cucullaea* I, *Cucullaea* II and the Submeseta members). This distinction between the lower three, and the upper three members of the La Meseta Formation is in line with the suggestion of Porebski (2000) that this unconformable boundary represents a major eustatic sea level low at 49.5 Ma. The continued shallowing upward trend following this proposed eustatic low is more likely to have been related to localised changes in relative sea level within the palaeovalley, probably as a consequence of infilling of the incised valley estuary system, than to have been related to global eustatic changes.

The waters were of variable productivity, with considerable variation within members. Each member showed an increase in productivity as it shallowed upward, moving closer to the shoreline and further up the estuary valley during each period of

fill, suggesting that the shallower waters in the La Meseta estuary have greater nutrient levels than the deeper waters.

The dinoflagellate cyst assemblage living within this environment was a mixed peridinialean and gonyaulacalean, with the most numerous dinoflagellate cyst species being *Enneadocysta partridgei*. The abundance of *E. partridgei* related to lower relative sea level and high productivity indicates that the linking by Askin *et al.* (1991) of this species to conditions that would be found within an estuary, e.g. hypopycnal conditions, is correct. The flora also had a strong input from the acritarch taxon *Enigmadinium cyclindfloriferum*. The presence of this taxon is mostly observed within intervals identified as stressed conditions, possibly in a wave-dominated, outer part of the estuary, given the strong marine influence.

The species richness data can be interpreted to show a record of the incision, deepening and sediment infilling represented by each individual member throughout the La Meseta Formation. The absolute abundance data suggests a correlation between high levels of absolute abundance and low levels of productivity and species richness, indicative of cyclical intervals of stressed environments.

The three formations studies are all shallow marine, near-shore settings, but the differences in the floral compositions and their changes through the sections reflects the importance of the depositional environment, e.g. tectonism, sediment supply/accommodations space, eustatic change and the properties of the individual water masses.

Chapter 5- Environmental Interpretation and Analysis Of Terrestrial Floras

5.1 - QUANTITATIVE ANALYSIS

5.1.1 - López de Bertodano Formation

The terrestrial microfloral assemblage in the López de Bertodano Formation is characterised by an overall stable assemblage with a few notable changes in relative abundance (Fig. 5.1). Pollen derived from the *Podocarpus* conifer is the most dominant, individual pollen taxon, comprising between 16 to 30% of the assemblage throughout the part of the López de Bertodano Formation present. Other gymnosperm pollen present (including *P. mawsonii*) represent 2 to 17% of the microfloral assemblage. Cryptogam spores are also common throughout the section, representing 21 to 68% of the microflora and consist primarily of *Cyathidites* sp. and *Baculatisporites* sp. Pollen from the angiosperm component comprise 7 to 52% of the assemblage, with *Nothofagus* being the largest individual pollen producer within this group.

Other selected ecologically-significant minor groups present within the López de Bertodano Formation include: the fern *Azolla* (Salviniaceae), which has a small but consistent presence throughout the section; the conifer *Microcachrys*, which has an infrequent and low abundance; the angiosperm families Casuarinaceae and Palmae, both showing occasional, infrequent occurrences predominantly in the lower part of the López de Bertodano Unit 10 section.

Trends seen within the López de Bertodano Formation include an overall slight decline in the conifer relative abundance. This is due primarily to a decline

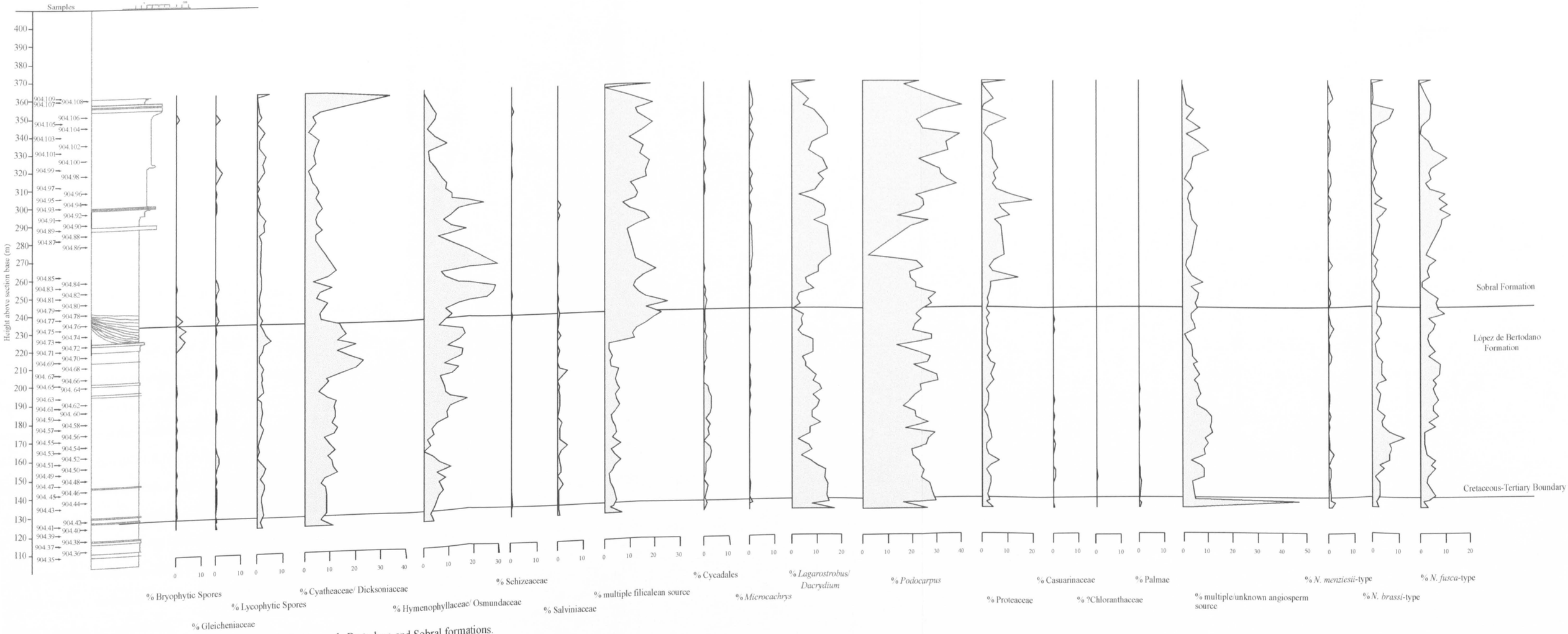


Fig 5.1. Stratigraphic variation of spores and pollen taxa in the Lopez de Bertodano and Sobral formations.

in the proportion of *P. mawsonii* pollen towards the top of the formation. Other changes apparent within the assemblages include a change in the distribution and relative abundance of some of the fern families. The most distinct changes seen occur in the Cyatheaceae/Dicksoniaceae tree fern group and the multiple filicalean origin group. The tree ferns display a consistent relative abundance through much of the formation of ~10%, before suddenly declining in abundance to ~5% (DJ 904.63) and then rising to an abundance high of 23% (DJ 904.68). Abundance then declines in the uppermost horizons towards the top of the formation. This group appears to show an inverse relationship with *P. mawsonii* pollen of the *Lagarostrobus* conifer. The multiple filicalean source group is a minor component throughout much of the López de Bertodano Formation, comprising ~4% of the flora. In the uppermost section (DJ 904.70-72) abundance increases to 12% and continues to increase across the López de Bertodano and Sobral boundary.

The majority of the spore and pollen taxa identified are associated with a forest, dominated by podocarp conifers in the canopy and a secondary component of Proteaceae, with an understorey of tree-ferns and shrubby Proteaceae angiosperms. The decline in the relative abundance of pollen (*Phyllocladidites mawsonii*) produced by the conifer *L. franklinii* suggests a decline in the annual rainfall rate. *Lagarostrobus franklinii* requires an annual rainfall rate of 1200mm (see above) and its decline would indicate a drying-out of the rainforest. A drying-out of the forest could also explain the changes among the ferns. A reduction of the conifer component would cause further openings in the canopy allowing greater development of the fern understorey (including the tree ferns), or the expansion of a fringing fern community.

The presence of the Casuarinaceae and the Palmae indicate a more warm

temperate aspect to the rainforest, although given the low abundance of these taxa, this is likely to have been small and probably localised in areas of low altitude and sheltered aspect. These taxa are likely to have been existing towards the limit of their climatic tolerance. Their presence is thought likely to be representative of frost-free conditions existing on parts of the peninsula landmass (Askin, 1994) The presence of the aquatic fern *Azolla* (Salviniaceae) is indicative of the presence of open freshwater bodies, as is the presence of the bryophytes (sphagnum mosses). These may have existed on coastal plain habitats. The low abundance of the conifer *Microcachrys* suggests the additional presence of a biogeographically limited, cold, high altitude setting (Askin, 1990).

5.1.2 - Sobral Formation

The microfloral assemblage of the Sobral Formation is characterised by the presence of the same major taxonomic groups as were present in the underlying López de Bertodano Formation (Fig. 5.1). Differences between the two formations arise when comparing the proportions of the groups present and the stratigraphic variation they display. Ferns are responsible for the highest relative abundance, with 38 to 66% of the assemblage. The largest groups contributing to this are the multiple filicalean source group and the Hymenophyllaceae/Osmundaceae (the filmy ferns). The most dominant individual taxon present is again the *Podocarpus* conifer, comprising between 2 to 39% of the flora. *Lagorostrobus* again forms a secondary part of the conifer component contributing up to ~15% of the total assemblage. The angiosperm group represents up to 35% of the flora, the majority of which is due to contributions from *Nothofagus*.

Minor ecologically-significant taxonomic groups present within the Sobral

Formation include: the Salviniaceae, which shows a lower abundance and less frequent occurrence than in the López de Bertodano Formation; *Microcachrys*, which displays a higher abundance and more frequent stratigraphic occurrence than seen in the lower López de Bertodano Formation; the Casuarinaceae have a single isolated occurrence in the lowermost Sobral Formation.

Trends within the Sobral Formation include some which were observed to initiate in the uppermost López de Bertodano Formation. The decline in *P. mawsonii* seen above continues across the boundary into the lowermost Sobral Formation, reaching a relative abundance low (DJ 904.75) before increasing to higher levels (DJ 904.79-109). Abundance remains high through the rest of the Sobral section. Other trends continuing into the basal Sobral Formation include the decline in the relative abundance of the tree ferns (Cyatheaceae/Dicksoniaceae). This decline continues through the Sobral Formation (DJ 904.73-106), where it is the inverse of the increase in abundance of *P. mawsonii*, until abundance returns to higher levels in the uppermost sediments (DJ 904.107-9). The increase in relative abundance multiple filicalean source group seen in the uppermost López de Bertodano Formation reaches a peak (DJ 904.77) and remains at a high level through the rest of the Sobral Formation.

Trends occurring solely within the Sobral Formation are seen in the *Podocarpus* pollen abundance results. Relative abundance remains high in the lower part of the Sobral Formation, but then declines sharply (DJ 904.85). It quickly recovers, with an increasing trend (DJ 904.86-94) reaching a high of 39%. Abundance remains high after this, although a high amount of variation is observed. Other trends include the Hymenophyllaceae/Osmundaceae 'filmy fern' component. This group, although a notable component of the total fern

assemblage, displayed no remarkable trends in the underlying López de Bertodano Formation. In the lower Sobral Formation, it shows a series of sporadic abundance peaks reaching abundances of ~30%. Abundance declines from DJ 904.93 and the group disappears in the uppermost part of the section. The angiosperm family Proteaceae also shows a change in distribution in the Sobral Formation. In the lowermost Sobral section it shows a low abundance, low variability distribution. From DJ 904.81 upwards, the Proteaceae show peaks of higher abundance with a higher background relative abundance. It also displays higher variability. The bryophyte group, predominantly represented by sphagnum mosses, shows a distinct increase in abundance in the lower Sobral Formation.

Interpretation of the Sobral Formation terrestrial microflora indicates an increase in the annual rainfall rate in the lower part of the section, as seen by the increase in *P. mawsonii*. These higher levels of rainfall appear to remain consistent throughout the rest of the Sobral Formation present. The structure of the rainforest also undergoes changes: the changing proportions of the dominant *Podocarpus* conifer coincides with an increase in the relative abundance of the broad-leaved angiosperm Proteaceae. This combined with the increase in the 'filmy fern' component and the multiple filicalean source group suggest an increase in the dominance of the rainforest understorey and groundcover. This change could be due to an increase in volcanism (see Chapter 4) and subsequent disturbance. The apparent inverse relationship seen between *P. mawsonii* and the tree ferns suggests competition between the two groups in the rainforest canopy/understorey controlled by annual rainfall

Among the less abundant taxa, the disappearance of the Palmae and the Casuarinaceae indicates the loss of the frost-free conditions seen in the López de

Bertodano Formation and the corresponding increase in *Microcachrys* suggests an expansion in the cold, high altitude flora. This increase could be due either to an expansion in the altitudinal zones of the flora caused by a fall in mean annual air temperature (MAT) or to increased run-off from the high altitude areas as a consequence of increased precipitation. The increase in the bryophyte group mentioned above seems to correlate with the interval of time during which incision of the basal channel unit occurred in the lowermost Sobral Formation. This change in the local environment could provide new space for the mosses to grow in suitably damp conditions, as well as being more likely to be transported and preserved in the basin.

5.1.3 - Cross Valley Formation

The terrestrial microflora of the Paleocene Cross Valley Formation spot samples is patchy, with poor preservation in DJ 912.1-2 and DJ 913. Only DJ 908 contains a more representative terrestrial microfloral assemblage (Fig. 5.2).

The most notable feature of the Cross Valley Formation samples is the large proportion of spores present in the assemblages, as with the Sobral Formation (see above) this could suggest a disturbed environment. In two of the four samples (DJ 912.2 and DJ 908), spores are dominant, in DJ 912.1 they have a parity of relative abundance with the gymnosperm component, and in DJ 913 they form a major part of the flora, behind the gymnosperms.

DJ 908 contains a terrestrial microflora that is more comparable to that seen in the López de Bertodano and Sobral formations in both the proportion of the groups and taxa present. The largest group present is the spores, represented primarily by the 'filmy fern' spore *Baculatisporites comaumensis* (7.77%) and the

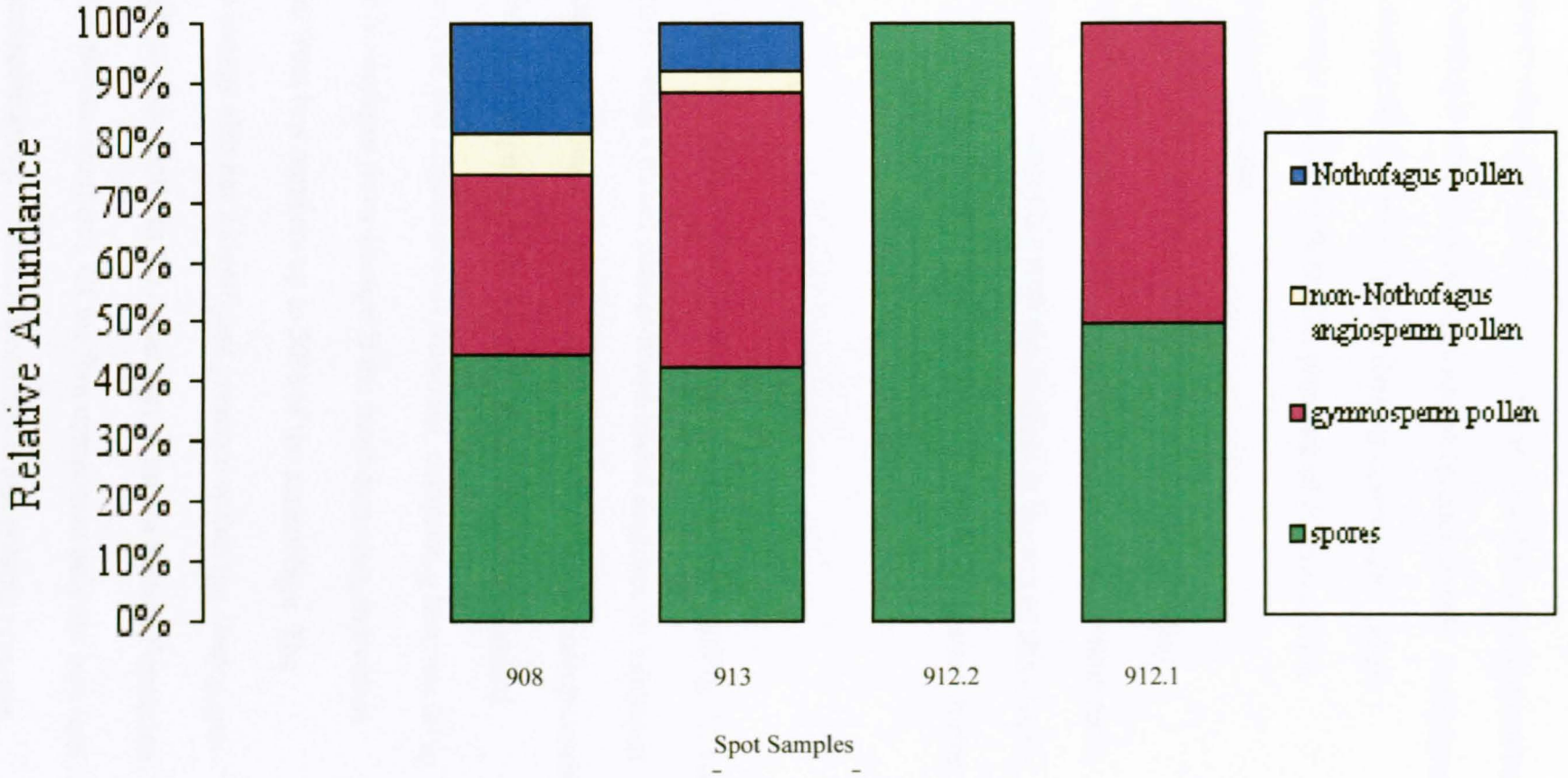


Fig. 5.2 Distribution of miospores in the Cross Valley Formation

tree fern *Cyathidites* spp. (6.66%). The largest individual taxon present is the gymnosperm *Podocarpidites* spp. (24.66%). The microflora that is present can be described as representing a mixed conifer-angiosperm (predominantly *Nothofagus*) forest, with a substantial fern understorey and fringing community. Other angiosperms are a minor part of the flora. The presence of *P. mawsonii* is suggestive of high annual rainfall.

Little more information can be ascertained from the Cross Valley Formation spot samples due to the paucity of information contained within them. What can be ascertained is compatible with the findings in the rest of this chapter for the terrestrial flora and conditions on the Antarctic Peninsula landmass during the Paleocene.

5.1.4 - La Meseta Formation

The terrestrial microflora of the Eocene La Meseta Formation (Fig. 5.3) is characterised as representing a mixed conifer-broad-leafed angiosperm rainforest. It has an overall stable assemblage with few significant changes in the composition of the flora. Podocarp conifer pollen is the largest group present, represented primarily by *Podocarpus* and *Lagarostrobus franklinii*, comprising between 30 to 60%. However, the *Nothofagus fusca*-lineage is the most dominant, individual taxon, and is present from low numbers up to 50% of the assemblage. The *Nothofagus brassii*-lineage also has a significant presence within the *Nothofagus* component, although predominantly in the upper part of the La Meseta Formation (*Cucullaea* II and Submeseta members). Of the fern component only the tree fern Cyatheaceae/Dicksoniaceae group maintains a consistent and notable presence throughout the La Meseta Formation.

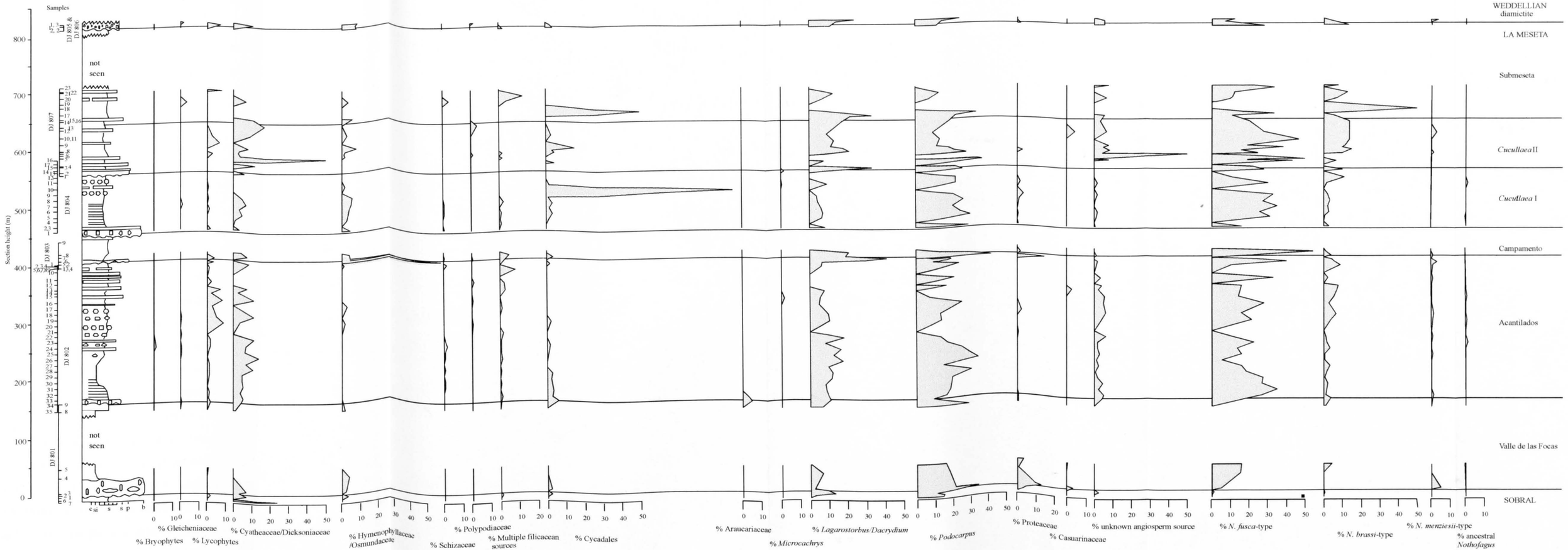


Fig 5.3. Stratigraphic variation of spores and pollen in the La Meseta Formation.

Other minor groups present with an ecological significance include: the conifer *Microcachrys*, which is represented only by a few sporadic and low abundance occurrences; the broad-leaved angiosperm Proteaceae, which have a notable presence in the lowermost La Meseta Formation (Valle de las Focas Member) but have a erratic and low abundance presence in the rest of the formation; the Casuarinaceae which have a very low abundance and infrequent distribution.

Trends seen within the La Meseta Formation are subtle. The major pollen taxa present representing: *N. fusca*-type, *Podocarpidites* spp., *P. mawsonii* and *N. brassii*-type, have nearly identical stratigraphic distributions and similar patterns of relative abundance. The tree ferns, Cyatheaceae/Dicksoniaceae, also have a similar pattern of stratigraphic distribution, although their relative abundance is markedly less. The pollen *P. mawsonii* and *N. brassii*-type both show a slight increase in relative abundance in the upper part of the La Meseta Formation (*Cucullaea* I to Submeseta members). The slight increase in abundance in the *N. brassii*-group in the upper part of the La Meseta Formation seen in Fig. 5.3 suggests a possible decline in temperature in the upper part of the formation. This increase in the *N. brassii*-lineage is seen more sharply in Fig. 5.4. Here, when plotted as a proportion of the *Nothofagus* genus, it is observed to increase markedly in the upper three members of the La Meseta Formation. The *N. fusca* lineage shows its highest abundance as a proportion of the total *Nothofagus* assemblage (Fig. 5.4) in the lower part of the Acantilados Member. The fern component does not appear to show any trends through the La Meseta sequence (Fig. 5.3). The families present do occasionally show single individual abundance peaks, often positioned on or near member boundaries. These may relate to short-

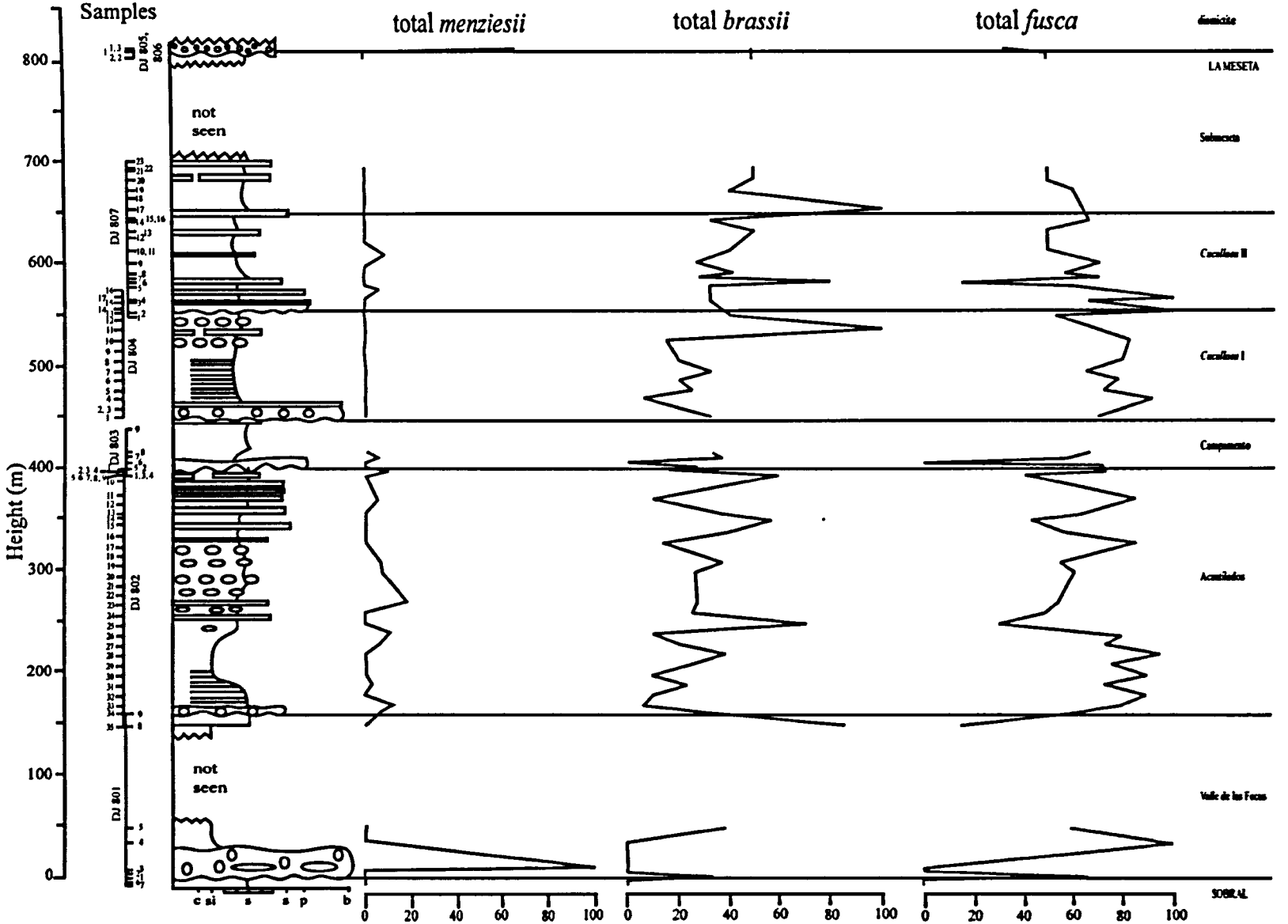


Fig. 5.4. *N. menziesii*, *N. brassii* and *N. fusca* lineages as a % of the total *Nothofagus* assemblage

lived, localised conditions around the margins of the palaeoestuarine margins that encouraged the growth of these floral assemblages, or alternately to the cyclical nature of sedimentation in the palaeovalley.

Occurrences of the minor groups mentioned previously are too low and sporadic to be able to pick out any trends occurring within these groups.

Interpretation of the La Meseta Formation terrestrial microflora indicates the presence of a mixed conifer-angiosperm (*Nothofagus*) rainforest. The consistent presence of *Lagarostrobus* pollen (*P. mawsonii*) indicates a high rainfall (1200mm/year) throughout the section (see above), with a slight increase in rainfall in the upper part of the La Meseta Formation. The higher proportion of the *N. fusca* lineage in the lower Acantilados Member suggests higher temperatures in the lower part of the La Meseta Formation (Pocknall, 1989). The increasing proportion of the *N. brassii* lineage suggests a cooling trend throughout the upper part of the La Meseta Formation (Pocknall, 1989). The sporadic occurrences of both the conifer *Microcachrys* and the angiosperm Casuarinaceae indicate the existence of both a cold, high altitude floral assemblage and a lowland tropical element within the floral population of the Antarctic Peninsula at this time.

5.2 - SEMI QUANTITATIVE ANALYSIS

5.2.1 - López de Bertodano Formation

The plant community associations (Fig. 5.5) show few changes through the López de Bertodano Formation section studied. The floral community of taxa assigned to plant associations is shown to be dominated by plants associated with a multistratal conifer-dominated rainforest environment. Minor changes are seen within the relative abundance of the multistratal rainforest group, as a slight

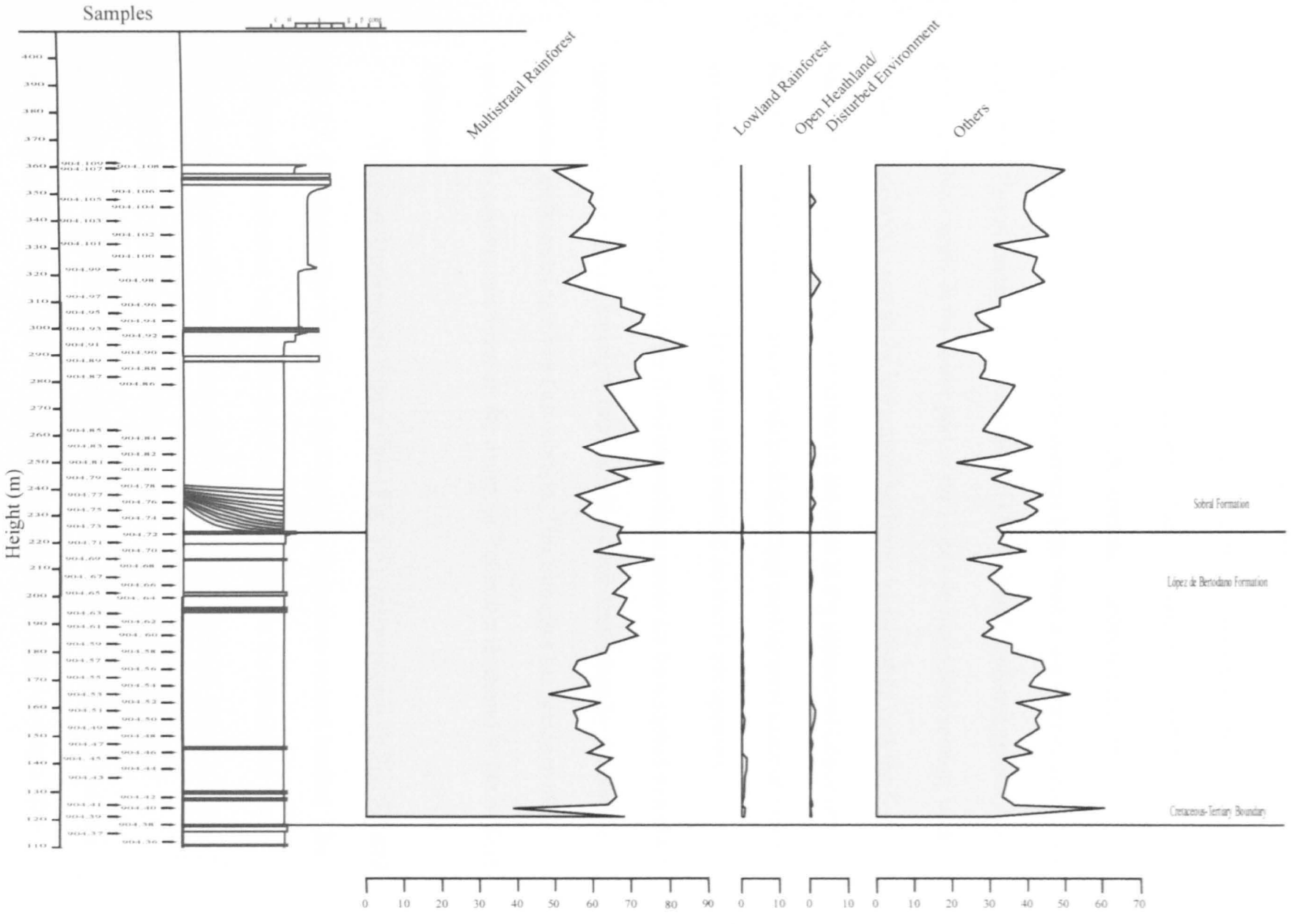


Fig. 5.5 López de Bertodano and Sobral Formation Plant Community Associations

decrease in relative abundance is seen in the lower section (DJ 904.42-57) followed by an increase in abundance leading to thickness of consistent higher abundance in the uppermost López de Bertodano Formation (DJ 904.70-2).

The two other groups present, the lowland rainforest and open heathland/disturbed environment associations both show a low relative abundance, but have different stratigraphic occurrences. The lowland rainforest association occurs predominantly in the lower part of the López de Bertodano section, with the highest abundances seen in the lowermost horizons, immediately post the K-T boundary. Rare and sporadic occurrences are seen in the uppermost López de Bertodano Formation. The open heathland/disturbed environment association has sporadic occurrences seen throughout the section; no trends are apparent.

A significant part of the floral assemblage could not be assigned to a plant community association, however, there is still a clear dominance by the conifer-dominated multistratal rainforest association. This indicates the presence of a widespread complex rainforest on the Antarctic Peninsula landmass in the earliest Paleocene.

The plant temperature regime data (Fig. 5.6), in common with Fig. 5.1 and Fig. 5.5, also shows few changes in the López de Bertodano section studied. The dominant temperature regime for the plants represented in the microflora is a mixed microthermal-mesothermal community. This community shows a slight decline through the measured section, but remains dominant. The mixed mesothermal-megathermal community has a relatively low abundance, but also displays a consistent presence throughout the section with little variation. The microthermal community association is minor, with very low abundances and only occasional sporadic occurrences.

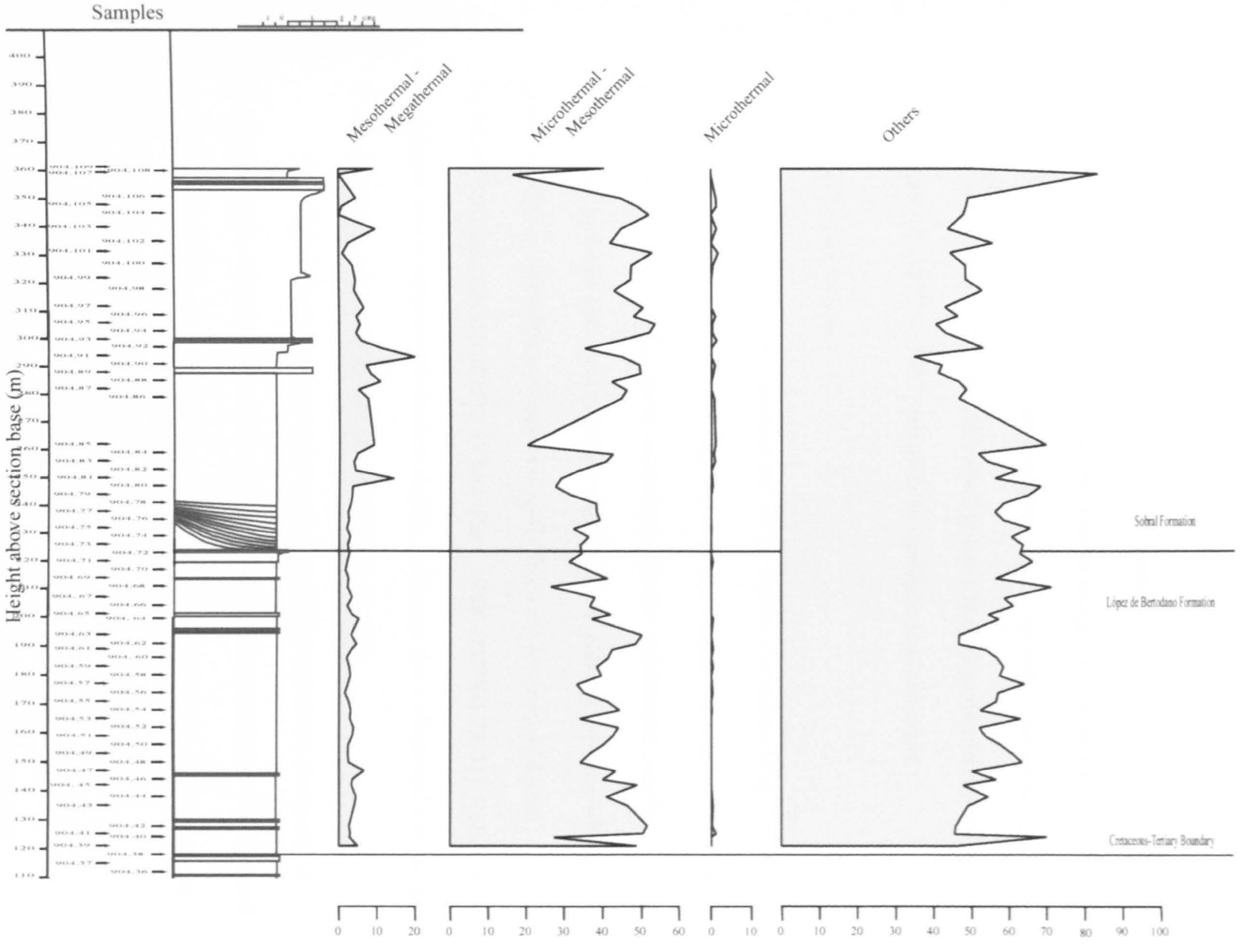


Fig. 5.6 López de Bertodano and Sobral formations Plant Temperature Regimes

The majority of the microflora are not assigned to a plant temperature regime due to both uncertainty in the identification of the parent taxa and the widespread habitats of others. The clear dominance of the microthermal-mesothermal group among those taxa that can be assigned, reveals a large part of the flora living in mean annual temperatures of 12°C to 14°C. The consistent presence of the mesothermal-megathermal group suggests a community living in temperatures of >20°C to 24°C, probably in sheltered lowland areas.

5.2.2 - Sobral Formation

The plant community associations of the Sobral Formation (Fig. 5.5) are very similar to those of the underlying López de Bertodano Formation. The floral community is dominated by plants associated with the multistratal conifer-dominated rainforest association. Minor changes in abundance are seen within this group, as relative abundance increases slightly from the lowermost Sobral Formation towards the top of Unit 1 (Macellari, 1988) (DJ 904.73-91). Relative abundance is then observed to decline slightly, although the multistratal rainforest is still dominant, to the top of the section (DJ 904.91-109).

The other minor groups present in the Sobral Formation are the lowland rainforest and the open heathland/disturbed environment associations. The lowland rainforest community has a single occurrence of low abundance in the lowermost Sobral Formation and is absent for the rest of the section. The open heathland/disturbed environment has a similar distribution to that seen in the López de Bertodano Formation (see above), with sporadic occurrences seen throughout the section and no apparent trends present.

The Sobral Formation microflora reveals a plant community still

predominantly a complex conifer-dominated multistratal rainforest. The disappearance of the lowland rainforest component indicates a change in conditions, possibly a change in the regional climate leading to a change in altitude of the floral zones, causing the lowland rainforest to become 'squeezed out'. The sporadic presence of the open heathland/disturbed environment could be as a consequence of landslides on the Antarctic Peninsula volcanic arc causing the temporary presence of open disturbed land suitable for colonisation.

The plant temperature regime data (Fig. 5.6) does show a change in the pattern of distribution from that seen in the López de Bertodano Formation. The mixed microthermal-mesothermal community is still the dominant part of the flora that has been assigned to a temperature regime. Having displayed a declining trend in abundance in the underlying formation (see above) this trend continues into the lower part of the Sobral Formation, reaching a low in sample (DJ 904.85). From this point, relative abundance increases to the high levels seen in the basal Paleocene, immediately post the K/T boundary event.

The change in distribution described above is also seen in both the mesothermal-megathermal and microthermal temperature regime groups. Both show an increase in relative abundance in the lower Sobral Formation (DJ 904.81-3). The mesothermal-megathermal group declines, with fluctuations, through to the top of the section. The microthermal group remains at a more consistent level of elevated abundance relative to its distribution in the López de Bertodano Formation. This would indicate an increase in both warm-loving plants and cold-loving plants at the same time, which may be possible if there were an increase in seasonality.

The increase in the relative abundance of the conifer *Lagarostrobus*,

coincident with the increase in these two groups suggests a general change in the climate of the region at this time. This increase in rainfall (see above) would lead to an increase in run-off. This would cause an increase in the representation of microflora from the hinterland, e.g. the microthermal community. The implication of this is that the increase in the abundance of the microthermal community may not have necessarily been caused by a decline in mean annual temperature. However, the disappearance of the Casuarinaceae and the Palmae, both seen in the lower section (see above), suggest that there may have been a decline in temperature during the Sobral Formation interval, as the frost-free conditions in the lowland areas of the Peninsula disappeared (Askin, 1994). A decline in temperature during this interval has been suggested in Dingle & Lavelle (1998), representing a mid-Paleocene temperature low. An initial increase in temperature is also possible with the increase shown in the mesothermal-megathermal group, although this then declines indicating a drop in MAT. However, it should also be noted that this increase is caused solely by an increase in the Proteaceae, an increase that may also have been caused by changes in the forest structure.

The climate is suggested to have moved towards a wetter state, with higher levels of rainfall during the time interval represented by the lower Sobral Formation. The majority of the flora that had a recognised temperature preference would have lived in temperatures of 12°C to 14°C, although there are indications of a decline in MAT in the upper part of the represented Sobral Formation.

5.2.3 - Cross Valley Formation

The plant community associations show the vast majority (49.7%) of the microflora, of the only representative sample among the Cross Valley Formation spot

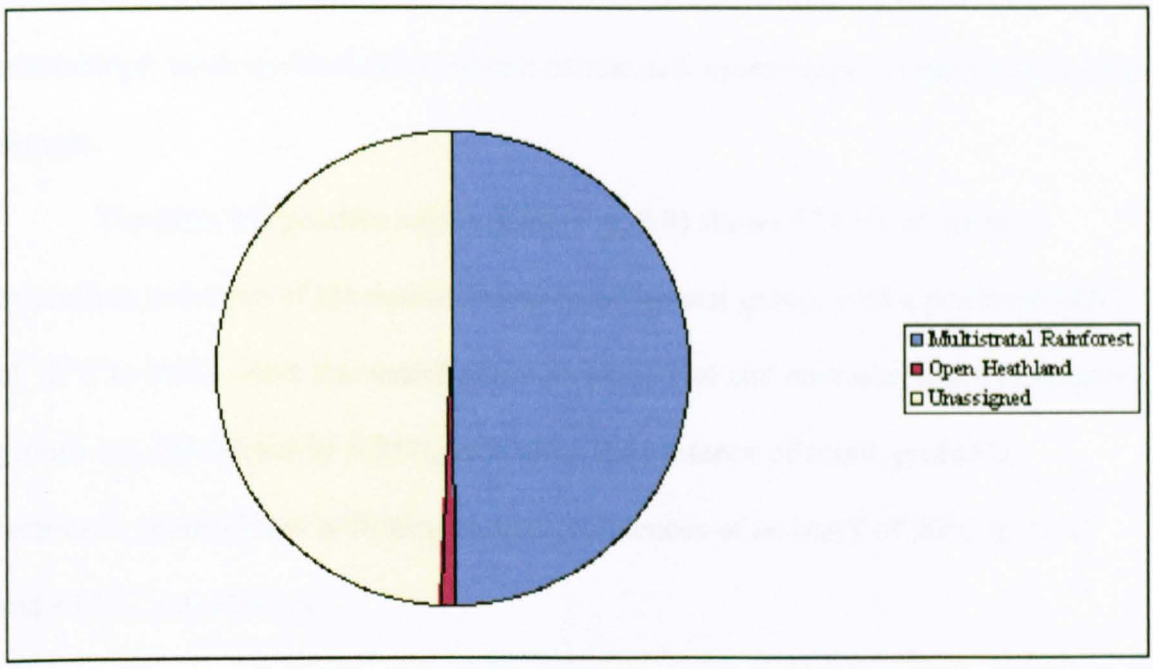


Fig. 5.7 Plant Community structure of the Cross Valley Formation

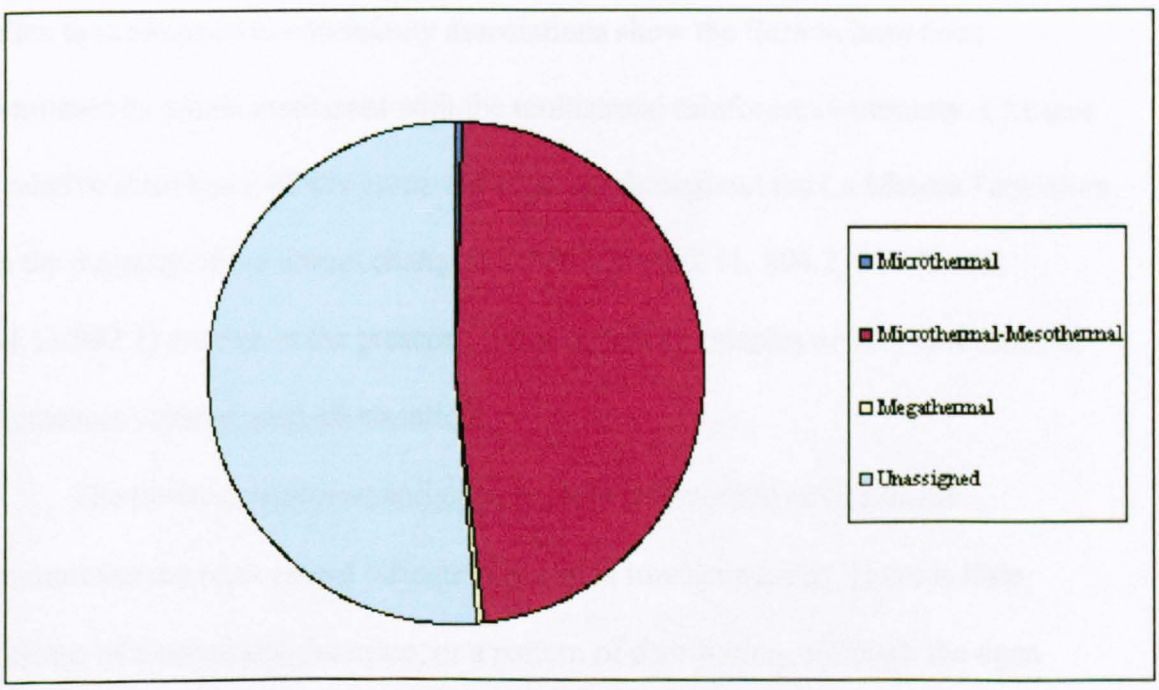


Fig. 5.8 Plant Temperature Regime of the Cross Valley Formation

samples, to be part of the multistratal rainforest community (Fig. 5.7). The open heathland/disturbed environment community is represented by 1% of the total assemblage, while no lowland rainforest community assemblage is represented in this sample.

The plant temperature regime data (Fig. 5.8) shows 47.84% of the total microflora to be part of the microthermal-mesothermal group, with a preferred MAT of 12°C to 14°C. Both the mesothermal-megathermal and microthermal temperature groups are represented by 0.25%, indicating the presence of small, probably restricted, communities with temperature preferences of an MAT of 20°C to 24°C and <12°C, respectively.

5.2.4 - La Meseta Formation

The plant community associations (Fig. 5.9) show few changes in composition between the identified groups. The microfloral community of spore and pollen taxa assigned to community associations show the flora to have been dominated by plants associated with the multistratal rainforest community. Changes in relative abundance of this group are apparent throughout the La Meseta Formation, but the majority of the abrupt changes (DJ 802.22, 802.11, 804.2, 804.12 and 804.13/807.2) are due to the presence either of barren samples or very low absolute abundances causing unusual variation.

The lowland rainforest and open heathland/disturbed environment communities are represented infrequently and in low abundance. There is little evidence of a consistent presence, or a pattern of distribution, although the open heathland/disturbed environment community does show a more consistent

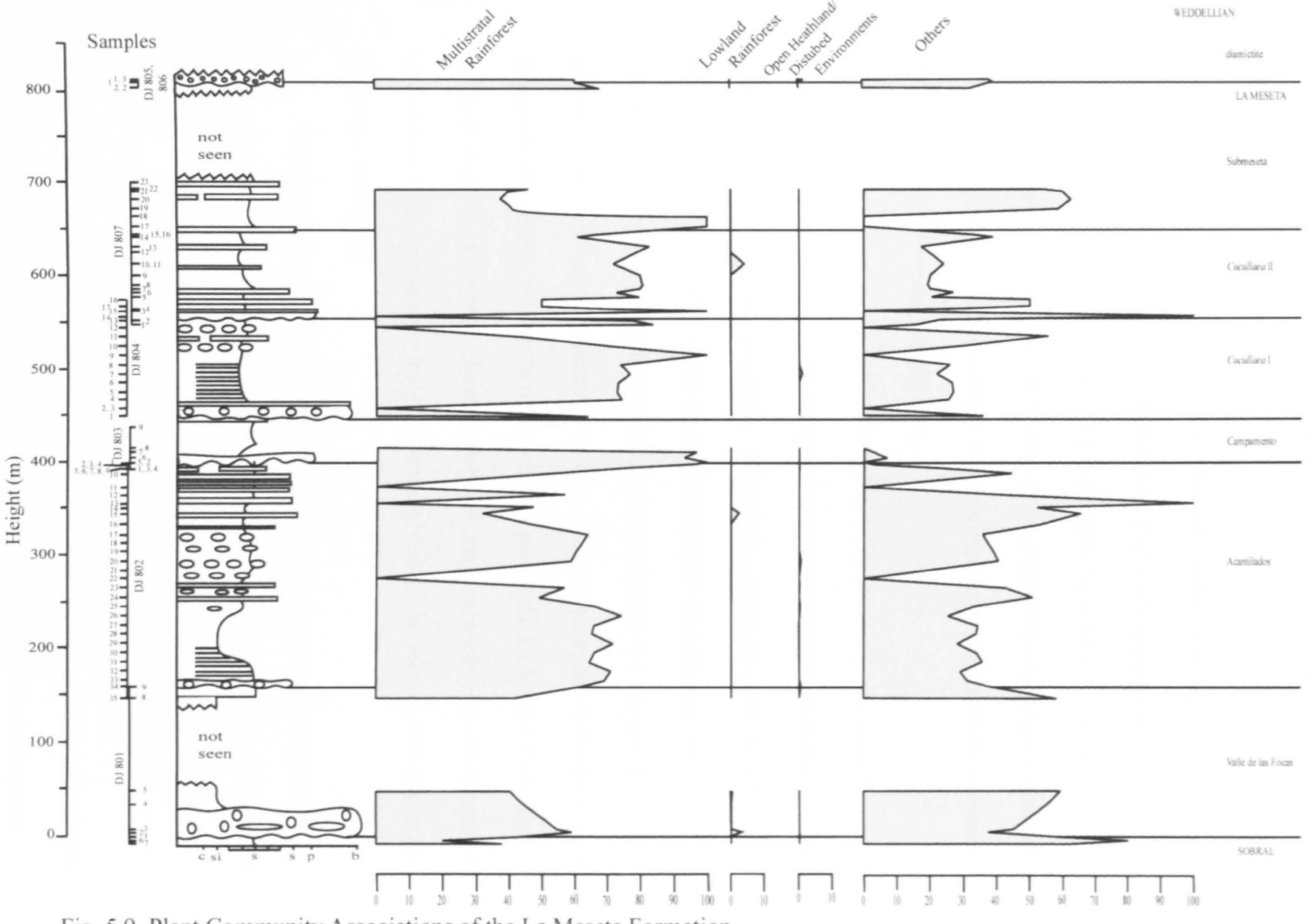


Fig. 5.9. Plant Community Associations of the La Meseta Formation

stratigraphic distribution, albeit with very low relative abundances, in the lower part of the Acantilados Member.

The dominance of the multistratal rainforest component suggest that, as with the Paleocene section (see above), the Antarctic Peninsula volcanic arc was covered in a complex multistratal rainforest. This consisted of a community of canopy trees (conifers and angiosperms), an understorey of tree ferns and other angiosperms and a ground cover of ferns. The understorey and groundcover elements would probably also have acted as fringing communities.

The plant temperature regime data (Fig. 5.10) shows that the dominant temperature regime for plants with a known recognised temperature preference is a microthermal-mesothermal community. It displays a dominance over the other two recognised plant temperature regimes, the mesothermal-megathermal and microthermal groups. The mesothermal-megathermal group has a frequent, though not consistent, distribution through the section. Its abundances are relatively low and its highest abundances are seen in the lowermost part of the La Meseta Formation (Valle de las Focas Member). The microthermal community has occasional, sporadic appearances of low abundance in the uppermost Valle de las Focas, Acantilados and *Cucullaea* I members.

The dominance of the microthermal-mesothermal community suggests that the majority of the palaeovegetation lived in temperatures of 12°C to 14°C, or a cool to warm temperate climate. The presence of the mesothermal-megathermal group shows the existence of a smaller group of taxa normally associated with warmer temperatures of 20°C to 24°C. These taxa are likely to have been living at the limits of their climatic tolerance and would only have been able to survive in sheltered areas. They are

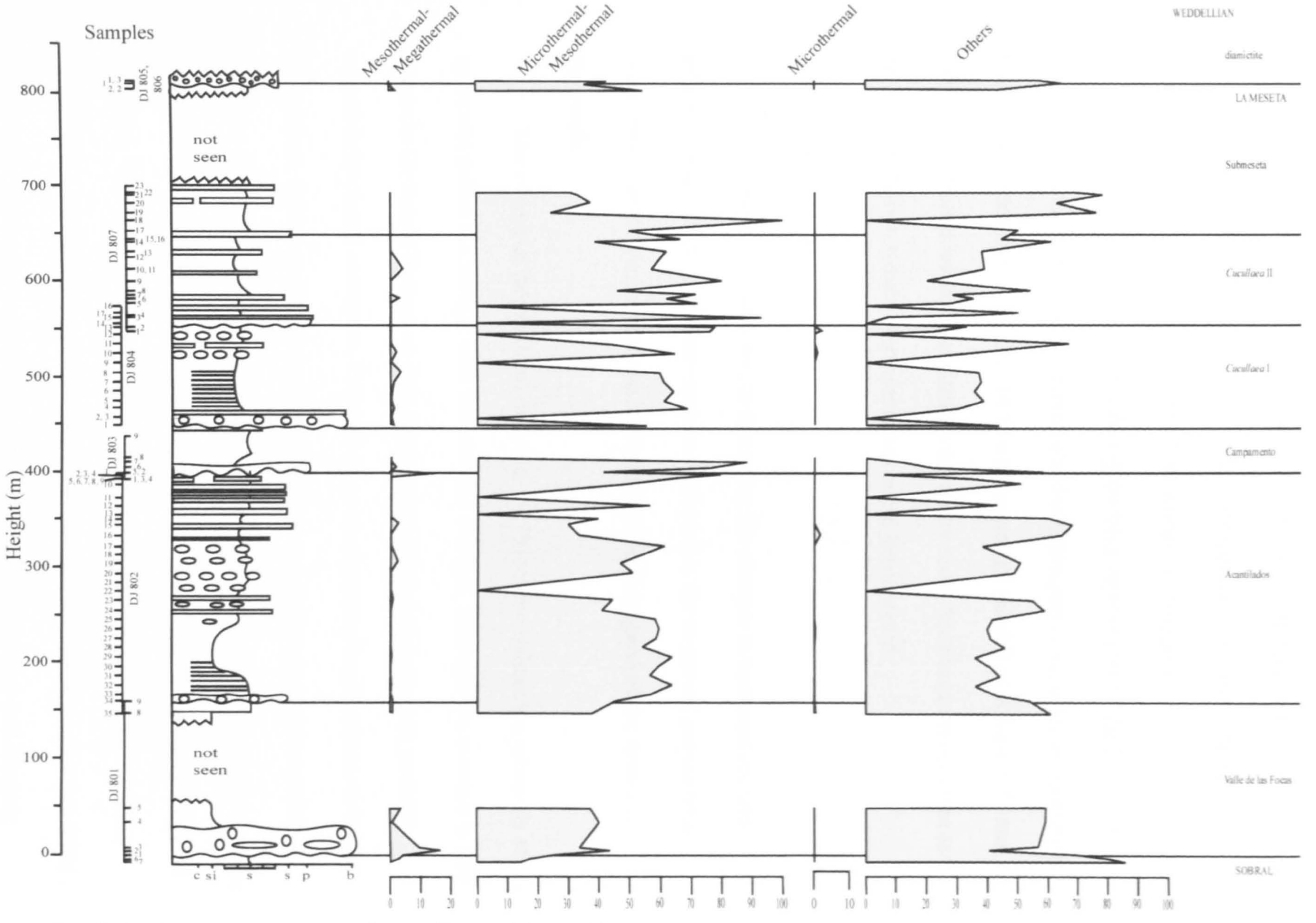


Fig. 5.10. Plant temperature regimes of the La Meseta Formation

unlikely to have been widespread. The microthermal group reveals the presence of a group living in temperatures of <12°C. These temperatures are likely to have been reached in the upper, hinterland areas of the Antarctic Peninsula.

Overall, the dominant floras during the time represented by the La Meseta Formation seems to have been a mixed conifer-angiosperm (*Nothofagus*) rainforest living in temperatures of 12°C to 14°C with an annual rainfall of at least 1200mm. Elements of both a lowland rainforest community and a heathland environment are present, but have a reduced representation in the fossil record.

5.3 – Conclusions

The aim of this study was to determine the changes in composition, and related changes in the palaeoenvironment reflected in the terrestrial palynofloras recovered from the shallow marine sediments of the Paleogene of the Antarctic Peninsula.

The terrestrial palynofloral record in the Paleocene indicates a relatively stable spore and pollen assemblage with few obvious changes, although the increase in fern spores in the Sobral Formation does suggest an unsettled environment, possibly related in increased volcanism. Pollen from podocarpaceous conifers dominated the assemblage. Other conifers (*Lagarostrobus*), fern spores and angiosperm pollen comprise the rest of the assemblage. The dominant plant community both through the López de Bertodano and the Sobral formations was a multistratal conifer-dominated rainforest. This was comprised predominantly by podocarpaceous conifers in the canopy and a secondary component of proteaceae, with an understorey of tree-ferns and shrubby proteaceae angiosperms. An additional ground cover and fringing community of ferns was also present. Smaller floral communities are also observed.

The dominant temperature regime for the plants represented in the microflora is a mixed microthermal-mesothermal community growing in an MAT of 12°C to 14°C and an annual rainfall of 1200mm.

Changes in the palaeoenvironment of the Antarctic Peninsula during the Paleocene are more apparent from the distribution of less abundant taxa and the smaller floral communities. The conifer *Lagarostrobus*, the distribution of which is linked to high levels of annual rainfall, increases in relative abundance in the lower Sobral Formation, following a decline in the upper López de Bertodano Formation. The disappearance of the Palmae and the Casuarinaceae in the upper López de Bertodano/lower Sobral Formation indicates the loss of a warmer frost-free influence seen in the lower López de Bertodano Formation. The corresponding increase in *Microcachrys* suggests an expansion in the cold, high altitude flora in the lower Sobral Formation.

The climate during the López de Bertodano Formation interval appears to have been relatively uniform, with an MAT of 12°C to 14°C and an annual rainfall of 1200mm. In the upper López de Bertodano Formation, conditions seem to have altered as levels of rainfall declined and the loss of the Palmae family, and decline of the Casuarinaceae, indicate a fall in temperature. The climate is then suggested to have moved towards a wetter state, with higher levels of rainfall during the time interval represented by the lower Sobral Formation. The majority of the flora that had a recognised temperature preference would have lived in temperatures of 12°C to 14°C, although there are indications of a decline in MAT in the upper part of the represented Sobral Formation.

No trends can be discerned through the Cross Valley Formation in either quantitative or semi-quantitative analysis due to the lack of a consistent sampled

section. However, the single sample that display a well-preserved flora shows conditions, both in the plant community composition and in the plant temperature regime, that are compatible with other results in this section.

The terrestrial palynofloral record in the Eocene (La Meseta Formation) indicates a relatively stable spore and pollen assemblage. There are several abrupt changes in the floral composition and these may relate to short-lived, localised conditions around the margins of the palaeoestuarine margins that encouraged the growth of these floral assemblages, or alternately, to the cyclical nature of sedimentation in the palaeovalley. There are some long-term trends apparent, noteably the increase in the proportion of the *N. brassi* group in the upper part of the formation and a similar increase in *P. mawsonii* pollen. A peak in the proportion of the *N. fusca* group in the lower Acantilados Member is also noted. Podocarp conifer pollen is the largest group present, represented primarily by *Podocarpus* and *Lagarostrobus franklinii*, although the *Nothofagus fusca*-lineage is the most dominant, individual taxon.

The dominant plant community was a mixed conifer-broad-leafed angiosperm (*Nothofagus*) multistratal rainforest. The structure of the forest would have been similar to that seen above, although the higher relative abundance of angiosperm pollen (*Nothofagus*) would ensure a larger presence of angiosperms within the forest canopy. The dominant temperature regimes for the plants represented in the microflora is a mixed microthermal-mesothermal community growing in an MAT of 12°C to 14°C, with an annual rainfall of 1200mm. This must be placed, however, against the quantitative evidence from the *Nothofagus* community for a decline in temperatures in the upper part of the formation, as well as a possible increase in annual rainfall, suggested by the increase in *P. mawsonii*.

The same quantitative evidence suggests a warm temperature peak in the lower Acantilados Member.

Overall, the dominant floras during the time represented by the La Meseta Formation seems to have been a mixed conifer-angiosperm (*Nothofagus*) rainforest, microthermal-mesothermal community living in an MAT of 12°C to 14°C with an annual rainfall of at least 1200mm.

CHAPTER 6 – CONCLUSIONS

The aims of the project were to provide a high-resolution palaeoenvironmental analysis of marine and terrestrial palynofloras from selected Paleocene and Eocene sections from Seymour Island. One hundred and fifty-nine samples from the Paleogene of the Antarctic Peninsula, supplied by the British Antarctic Survey, were processed and analysed by quantitative palynological methods (see Chapter 2). The majority of samples contained abundant and well-preserved palynofloras with little thermal alteration.

6.1 – PALEOCENE

6.1.1 - Marine

The marine palynofloral record in the Paleocene shows the palaeoenvironment of the López de Bertodano Formation to have been a shallow marine setting, with evidence for a regressive trend through much of the measured section There is also evidence for a corresponding transgression in the uppermost interval of the formation. The waters were moderately to highly productive, with an apparently terrestrial source for the nutrients.

The dinoflagellate cyst assemblage living within this environment was predominantly peridinialean, although the lowermost interval contains a different assemblage of low abundance and species richness that contrasts with the rest of the section. This is thought to represent a stressed or restricted environment. The overlying López de Bertodano Formation interval contains an assemblage dominated first by *P. pyrophorum* and *Spinidinium* spp., and then by *T. evittii.* This turnover represents an influx of warm water from lower latitudes (Brinkhuis *et al.*, 1998)

characterised by the low latitude species *T. evittii*. This event is also seen in Australian and New Zealand sections (Willumson, 2000). The dominance of *T. evittii* is not long-lived, and the assemblage reverts to its previous composition. The productivity levels of the waters do not seem to have been affected by the changes in temperature.

The Sobral Formation is separated from the underlying López de Bertodano Formation by a localised hiatus up to 100m and of apparently little duration in time (Askin, 1988). Shallow marine conditions are suggested; the lower part of the section is represented by the top of a transgressive cycle that began in the upper López de Bertodano Formation, while the rest of the measured section reflects a regression that continues through the top of the measured section. This regression has been interpreted as the development of a delta system (Macellari, 1988). The waters were consistently highly productive throughout the section present, indicating eutrophic conditions with little variation in nutrient availability.

The dinoflagellate cyst assemblage living within this environment was predominantly peridinialean, *P. pyrophorum* was the most abundant individual dinoflagellate cyst species present with *Spinidinium* spp. another important taxa. The shallow water acritarch genus *Micrhystridium* spp. (Wall, 1965) was also a major part of the flora during the transgressive sub-cycle in the lower part of the formation. Both species richness and abundance declined through the measured interval of the Sobral Formation, again reflecting the regressive trend discussed above.

6.1.2 - Terrestrial

The terrestrial palynofloral record in the Paleocene indicates a relatively stable spore and pollen assemblage with few obvious changes. Pollen from podocarpaceous

conifers dominated the assemblage. Other conifers (*Lagarostrobus*), fern spores and angiosperm pollen comprise the rest of the assemblage. The dominant plant community both through the López de Bertodano and the Sobral formations was a multistratal conifer-dominated rainforest. This was comprised predominantly by podocarpaceous conifers in the canopy and a secondary component of proteaceae, with an understorey of tree-ferns and shrubby proteaceae angiosperms. An additional ground cover and fringing community of ferns was also present. Smaller floral communities are also observed. The dominant temperature regime for the plants represented in the microflora is a mixed microthermal-mesothermal community growing in an MAT of 12°C to 14°C and an annual rainfall of 1200mm.

Changes in the palaeoenvironment of the Antarctic Peninsula during the Paleocene are more apparent from the distribution of less abundant taxa and the smaller floral communities. The conifer *Lagarostrobus*, the distribution of which is linked to high levels of annual rainfall, increases in relative abundance in the lower Sobral Formation, following a decline in the upper López de Bertodano Formation. The disappearance of the Palmae and the Casuarinaceae in the upper López de Bertodano/lower Sobral Formation indicates the loss of the frost free flora living in the lowlands of the peninsula seen in the lower López de Bertodano Formation. The corresponding increase in *Microcachrys* suggests an expansion in the cold, high altitude flora in the lower Sobral Formation. An increase in the number of fern spores in the Sobral Formation suggests an increase in disturbance, possibly linked to volcanism.

The climate during the López de Bertodano Formation interval appears to have been relatively uniform, with MAT of 12°C to 14°C and an annual rainfall of 1200mm. In the upper López de Bertodano Formation, conditions seem to have

altered as levels of rainfall declined and the loss of the Palmae family, and decline of the Casuarinaceae, indicate a fall in temperature. The climate is then suggested to have moved towards a wetter state, with higher levels of rainfall during the time interval represented by the lower Sobral Formation. The majority of the flora that had a recognised temperature preference would have lived in temperatures of 12°C to 14°C, although there are indications of a decline in MAT in the upper part of the represented Sobral Formation.

6.2 – Eocene

6.2.1 - Marine

The Eocene La Meseta Formation lies unconformably over the Paleocene Sobral Formation into which the younger sediments are deposited within an fault-bounded incised estuary valley. The samples from the La Meseta Formation show the palaeoenvironment to have been an overall shallow marine setting, with a further shallowing up trend present in the upper three members (*Cucullaea* I, *Cucullaea* II and the Submeseta members). This distinction between the lower three, and the upper three members of the La Meseta Formation is also in line with the suggestion of Porebski (2000) that this unconformable boundary represents a major eustatic sea level low at 49.5 Ma. The continued shallowing upward trend following this proposed eustatic low is more likely to have been related to localised changes in relative sea level within the palaeovalley, probably as a consequence of infilling of the incised valley estuary system, than to have been related to global eustatic changes.

The waters were of variable productivity, with considerable variation within members. Each member showed an increase in productivity as it shallowed upward, moving closer to the shoreline and further up the estuary valley during each period of

fill, suggesting that the shallower waters in the La Meseta estuary have greater nutrient levels than the deeper waters. This may have been due to embayment of shallower waters within the estuary, allowing the concentration of peridinioid dinoflagellate cysts in nutrient-rich waters. The deeper waters may have been open and of higher energy, causing the diffusion of the nutrient-load and creating conditions less favourable to the peridinioid dinoflagellate cyst community.

The dinoflagellate cyst assemblage living within this environment was a mixed peridinialean and gonyaulacalean, with the most numerous dinoflagellate cyst species being *Enneadocysta ursulae*. The abundance of *E. ursulae* related to lower relative sea level and high productivity indicates that the linking by Askin (1991) of this species to conditions that would be found within an estuary, e.g. hypopycnal conditions, is correct. The flora also had a strong input from the acritarch taxon *Enigmadinium cyclindfloriferum*. The presence of this taxon is mostly observed within intervals identified as stressed conditions, possibly in a wave-dominated, outer part of the estuary, given the strong marine influence.

The species richness data can be interpreted to show a record of the incision, deepening and sediment infilling represented by each individual member throughout the La Meseta Formation. The absolute abundance data suggests a correlation between high levels of absolute abundance and low levels of productivity and species richness, indicative of cyclical intervals of stressed environments. Both the absolute abundance data and species richness reach their highest values in the lower Acantilados Member (Fig. 4.6 C-D).

6.2.2 - Terrestrial

Interpretation of the La Meseta Formation terrestrial microflora reveals a

mixed conifer-broad-leafed angiosperm (*Nothofagus*) rainforest, with has an overall stable assemblage and few significant changes in the composition of the flora. Podocarp conifer pollen is the largest group present, represented primarily by *Podocarpus* and *Lagarostrobus franklinii*, although the *Nothofagus fusca*-lineage is the most dominant, individual taxon. The consistent presence of *Lagarostrobus* pollen (*P. mawsonii*) indicates a high rainfall (1200mm/year) throughout the section (see above). The sporadic occurrences of both the conifer *Microcachrys* and the angiosperm Casurinaceae indicate the existence of both a cold, high altitude floral assemblage and a lowland warmer element, respectively, within the floral population of the Antarctic Peninsula at this time. The high proportion of the *N. fusca*-lineage (Fig. 5.4) suggests that the warmest temperatures (Pocknall, 1989) are seen in the lower Acantilados Member. The increase of the *N. brassii*-lineage in the upper part of the formation, conversely indicates a cooling in the temperature.

The microfloral community of spore and pollen taxa assigned to community associations show the flora to have been dominated by plants associated with the multistratal rainforest community. The dominance of the multistratal rainforest component suggests that the Antarctic Peninsula volcanic arc was covered in a complex multistratal rainforest. This consisted of a community of canopy trees (conifers and angiosperms), an understorey of tree ferns and other angiosperms and a ground cover of ferns. The understorey and groundcover elements would probably also have acted as fringing communities.

Overall, the dominant floras during the time represented by the La Meseta Formation seems to have been a mixed conifer-angiosperm rainforest, microthermal-mesothermal community living in an MAT of 12°C to 14°C with an

annual rainfall of at least 1200mm. The warmest temperatures appear to have occurred in the lower Acantilados Member, while the coolest temperatures occurred during the deposition of the upper La Meseta Formation.

6.3 – Large scale Implications

6.3.1 - Sea Level Changes

The data generated for this study, when compared to published data relating to the global sea level curve (Fig. 6.1), show that, the sea level changes observed in this study are not comparable to the eustatic sea level curve of Haq *et al.* (1987). An exception in seen in the Eocene sequence boundary at 49.5 Ma (Porebski, 2000), which does correspond to the 49.5 Ma eustatic low of Haq *et al.* (1987) (Fig. 6.1A). All other ages are tentative. Another exception, though less obvious, is seen in the regressive trend in the upper López de Bertodano Formation and the Sobral Formation, despite the minor transgressive trend seen within it (Fig. 6.1B).

The data that does not correspond to the global eustatic curve (Haq *et al.*, 1987) is predominantly from the upper part of the La Meseta Formation. The transgressive trend apparent throughout this section (Fig. 6.1B) and which was also apparent to Wrenn & Hart (1988), is related to the infill of the La Meseta palaeovalley. The marine flora in this interval appears to be responding only to the local relative sea level changes affected by valley floor subsidence and sediment supply.

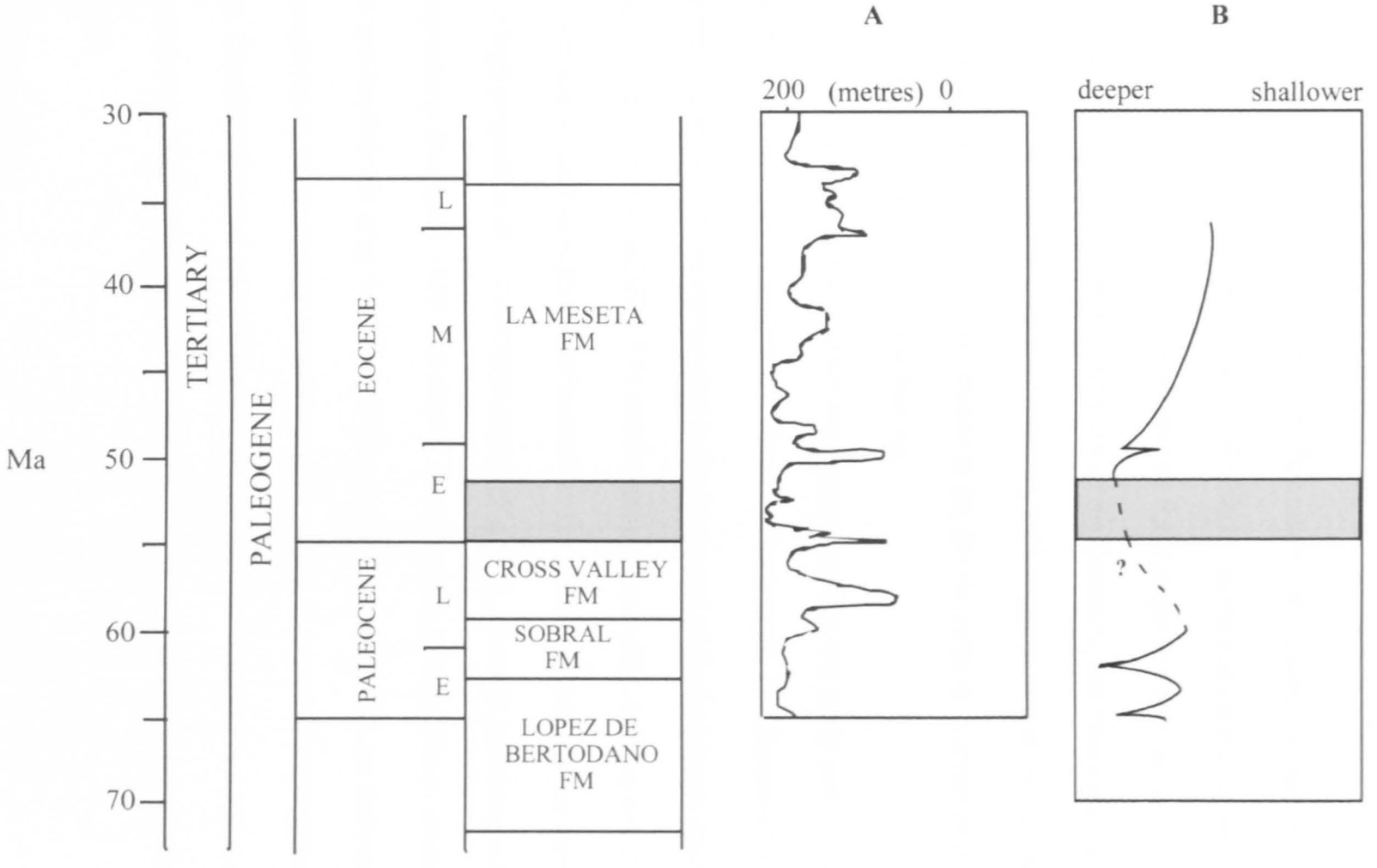


Fig. 6.1. Graph showing relative sea level changes during the Paleogene. Lithostratigraphy after Dingle & Lavelle (1998). A, Eustatic curve of Haq et al. (1987). B, Relative sea level change curves generated from this work.

6.3.2 - Temperature Changes

The results from this study show that relative temperature closely follows the temperature isotope curve of Zachos *et al.* (2001), where sufficient evidence is present. Fig. 6.2, a summary diagram, shows the changes that have occurred through the studied section. The warmest intervals of the Cenozoic, the Paleocene-Eocene Boundary and the Eocene Climatic Optimum, both occurred within the Paleogene. The global cooling event that culminated in the development of permanent ice sheets at the poles also began during this interval.

The Paleocene-Eocene event is not represented in this study and it is not clear that it is represented at all in Antarctic Peninsula sediments. The Eocene Climatic Optimum does seem to be represented in the lower part of the La Meseta Formation (Fig. 6.2). Both terrestrial (Fig. 5.4) and marine evidence (Fig. 4.5) indicate a warmer interval occurring during the deposition of the lower Acantilados. This is represented by a high proportion of *N. fusca*-type pollen (Fig. 5.4), a lineage preferring warmer temperatures (Pocknall, 1989) and also by dinoflagellate cyst maximum abundance and species richness (Fig. 4.6). The cooling event of the Mid and Late Eocene is represented by the increasing proportion of the cooler *N. brassii*-lineage pollen (Fig. 5.4). Dinoflagellate species richness and diversity is also observed to decline over this interval (Fig. 4.5) but it is unclear to what extent this may be due to more local factors.

6.4 – SUMMARY

The three Paleogene formations studied are all shallow marine, near-shore settings, but the differences in the marine floral compositions and their changes through the sections reflects the importance of the depositional environment, e.g.

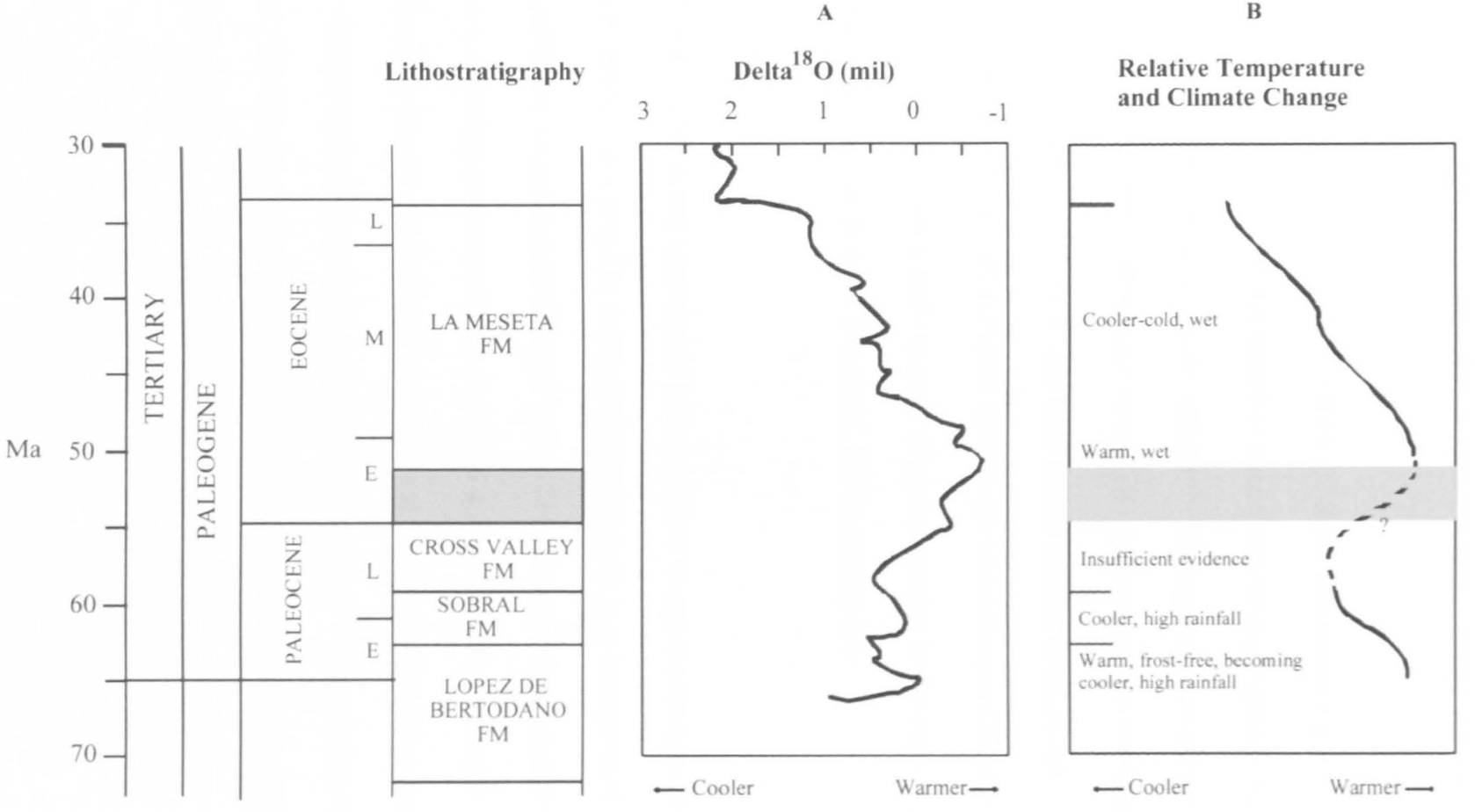


Fig. 6.2. Summary diagram for temperature and climate ghange during the Paleogene. Lithostratigraphy after Dingle & Lavelle (1998). A, isotope curve after Zachos *et al.* (2001). B, data generated in this study.

tectonism, sediment supply/accommodation space, eustatic change and the properties of the individual water masses.

The terrestrial palynomorph data from the three formations suggests the presence of a stable multistratal rainforest of mixed conifer-angiosperm dominance on the Antarctic Peninsula landmass throughout the Paleogene. The majority of the represented flora that grew within these forests, did so in conditions with a mean annual air temperature 12°C to 14°C and with annual rainfall in excess of 1200mm per year. Evidence indicates the presence of both the warm Eocene Climatic Optimum in the Early Eocene and the onset of global cooling in the upper La Meseta Formation, following the 49.5 Ma (Porebski, 2000) eustatic low.

6.5 - FUTURE WORK

Further work in this area should include efforts to age the sections more rigorously, using where possible multi-disciplinary teams, e.g. isotope work. This would enable the environmental changes to be placed in a tighter stratigraphic framework, allowing more correlation on a local and global scale. Research on establishing the history and dating of local tectonic events, and the effect that these have on the micro- and macroflora, as well as the pattern of sedimentation would also be useful. The effects of local events could then be more clearly distinguished from wider-scale events.

TAXONOMIC LISTING OF PALYNOMORPHS

List of palynomorphs with author citations

All palynomorph species are listed alphabetically, within their respective constituent groups. Presumed reworked forms are asterisked. Species illustrated are indicated in parentheses.

7.1 - Marine Species list

7.1.1 – Dinoflagellate Cysts

**Alisocysta* sp.

**Alisocysta circumtabulata* (Drugg) Stover & Evitt, 1977 (reworked in La Meseta Formation)

(Plate 1.1)

**Alisocysta reticulata* Damassa, 1979 (reworked in La Meseta Formation)

(Plate 1.2)

**Alisocysta rugolirata* Damassa 1979 (reworked in La Meseta Formation)

(Plate 1.3)

Arachnodinium antarcticum Wilson & Clowes, 1982

(Plate 1.4)

Batiacasphaera sp.

Batiacasphaera compta Drugg, 1970b

Batiacasphaera explanata (Bujak) Islam, 1983

Bourkidinium sp. indet.

**Bourkidinium cylinderatum* Dolding, 1992

(Plate 1.5)

**Bourkidinium psilatum* Ch Singh, 1983

Brigantedinium sp.

**Ceratiopsis striata* (Drugg) Lentin & Williams (reworked in La Meseta Formation)

(Plate 1.6)

Cerodinium sp. indet.

**Cerodinium speciosum* (Alberti) Lentin & Williams, 1987 (reworked in La Meseta Formation)

(Plate 1.7)

**Cometodinium whitei* (Deflandre & Courteville) Stover & Evitt, 1978

(Plate 2.1)

Cordosphaeridium minimum (Morgenroth) Benedek, 1972

(Plate 1.8)

**Cribroperidinium sp. indet.*

**Cribroperidinium edwardsii* (Cookson & Eisenack) Davey, 1969

(Plate 2.2)

**Cribroperidinium muderongense* (Cookson & Eisenack) Davey, 1969

Deflandrea sp. indet.

Deflandrea antarctica Wilson, 1967

(Plate 2.3)

**Deflandrea dartmooria* (Cookson & Eisenack) Lindgren, 1984) (reworked in La Meseta Formation)

**Deflandrea medcalfii* Stover, 1974 (reworked in La Meseta Formation)

(Plate 2.4)

Deflandrea phosphoritica Eisenack, 1938

**Diconodinium cristatum* Cookson & Eisenack emend. Morgan, 1977

*_Diconodinium multispinum_ (Deflandre & Cookson) Eisenack & Cookson emend. Morgan, 1977

(Plate 2.5)

*?_Dinogymnium_ sp.

*_Dinogymnium curvatum_ (Vozzhennikova) Lentin & Williams, 1973

Dino N Gen X of Askin 1988

Elytrocysta sp.

Enneadocysta partridgei Stover & Williams, 1995

(Plate 2.7)

The southern hemisphere form of _Areosphaeridium diktyoplokus_. A taxonomic re-evaluation of the _Areospheridium_ complex by Stover & Williams (1995) led to the reassignment of this taxon.

Enneadocysta partridgei sub-species A of Cocozza & Clarke, 1992

(Plate 2.8)

This sub-species has been similarly re-assigned.

*_Exochosphaeridium bifidum_ (Clarke & Verdier) Clarke _et al._, 1968

(Plate 3.1)

Hystrichokolpoma sp. indet

Hystrichosphaeridium sp. indet.

Hystrichosphaeridium astartes Sannemann, 1955

Hystrichosphaeridium compactum Davey& Williams, 1966

Hystrichosphaeridium cylinderatum Morgenroth, 1966

Hystrichosphaeridium latirictum Davey & Williams, 1966

*_Hystrichosphaeridium salpingophorum_ Deflandre emend. Davey & Williams, 1966

(Plate 3.2)

Hystrichosphaeridium truswelliae Wrenn & Hart, 1988

(Plate 3.3)

**Hystrichosphaeridium tubiferum* (Ehrenberg) Deflandre emend. Davey & Williams, 1966

(Plate 3.4)

Hystrichosphaeridium tubiferum brevispinium (Davey & Williams) Lentin & Williams, 1973

Impagidinium sp. indet.

Impagidinium maculatum (Cookson & Eisenack) Stover & Evitt, 1978

(Plate 3.5)

Impagidinium victorianum (Cookson & Eisenack) Stover & Evitt, 1978

(Plate 3.6)

Impletosphaeridium sp.

(Plate 3.7)

Isabelidinium sp.

**Isabelidinium pellucidum* (Deflandre & Cookson) Lentin & Williams, 1977

(reworked in La Meseta Formation)

Lejeunecysta sp. indet

Lejeunecysta fallax (Morgenroth) Artzner & Dörhöfer, 1978

(Plate 3.8)

Manumiella sp. indet

Manumiella druggii (Stover) Bujak & Davies, 1983

Microdinium sp.

(Plate 4.2)

**Nelsoniella aceras* Cookson & Eisenack, 1960

Octodinium askiniae Wrenn & Hart, 1988

(Plate 4.3)

**Odontochitina spinosa* Wilson, 1984

Oligosphaeridium sp.

**Oligosphaeridium complex* (White) Davey & Williams, 1966

**Oligosphaeridium patulum* Riding & Thomas, 1988

**Oligosphaeridium poculum* Jain, 1977

**Oligosphaeridium pulcherrimum* (Deflandre & Cookson) Davey & Williams, 1966

Operculodinium sp.

Operculodinium bergmannii (Archangelsky) Stover & Evitt, 1978

**Operculodinium radiculatum* Smith, 1992

Palaeocystodinium sp. indet.

**Palaeocystodinium australinium* (Cookson) Lentin & Williams, 1976 (reworked in La Meseta Formation)

Palaeocystodinium golzowense Alberti, 1961

(Plate 4.4)

**Palaeocystodinium granulatum* (Wilson) Lentin & Williams, 1976

**Palaeocystodinium lidiae* (Gorka) Davey, 1969

**Palaeoperidinium pyrophorum* Ehrenberg emend Sarjeant emend. Gocht & Wetzel, 1977 (reworked in La Meseta Formation)

(Plate 4.5)

Phelodinium sp.

**Phelodinium magnificum* (Stanley) Stover & Evitt

Phthanoperidinium sp.

Phthanoperidinium echinatum Eaton, 1976

**Pseudoceratium securigera* (Davey & Verdier) Bint

(Plate 5.1)

Pyxidiniopsis sp. indet.

Pyxidiniopsis crassimurata Wilson, 1988

(Plate 4.8)

Pyxidiniopsis delicata Wilson, 1988

(Plate 4.7)

Selenopemphix nephroides Benedek emend. Bujak, 1980

(Plate 5.2)

Senegalinium asymmetricum (Wilson) Stover & Evitt, 1978

Senegalinium dilwynense (Cookson & Eisenack) Stover & Evitt

(Plate 5.3)

Senegalinium obscurum (Drugg) Stover & Evitt

**Sepispinula ancoriferum* (Davey *et al.*) Lentin & Williams, 1993

Spinidinium sp. indet

Spinidinium bidwillii Cookson & Eisenack

Spinidinium colemanii Wrenn & Hart, 1988

(Plate 5.4)

**Spinidinium densispinatum* Stanley, 1965 (reworked in La Meseta Formation)

**Spinidinium essoi* Cookson & Eisenack, 1967 (reworked in La Meseta Formation)

(Plate 5.5)

**Spinidinium lanterna* Cookson & Eisenack, 1970 (reworked in La Meseta Formation)

Spinidinium luciae Wrenn & Hart, 1988

(Plate 5.6)

Spinidinium macmurdoense (Wilson) Lentin & Williams, 1976

(Plate 5.7)

**Spinidinium pilatum* (Stanley) Costa & Downie (reworked in La Meseta Formation)

(Plate 5.8)

Spinidinium sp. 1 of Askin, 1988

Spiniferites sp. indet

Spiniferites brevifurcatus Cookson & Eisenack, 1974

Spiniferites hyperacanthus (Deflandre & Cookson) Cookson & Eisenack, 1974

**Spiniferites microceras* (Cookson & Eisenack) Lentin & Williams, 1977 (reworked in La Meseta Formation)

(Plate 6.1)

**Spiniferites multibrevis* Cookson & Eisenack, 1974

Spiniferites pseudofurcatus (Klumpp) Sarjeant, 1970

(Plate 6.2)

Spiniferites ramosus (Ehrenberg) Loeblich & Loeblich, 1966

**Spiniferites ramosus brevifurcatus* (Cookson & Eisenack) Lentin & Williams, 1977

**Spiniferites ramosus reticulatus* (Davey & Williams) Lentin & Williams, 1973

Subtilisphaera sp.

(Plate 6.3)

**Systematophora ?palmula* Davey, 1982

Tanyosphaeridium sp.

**Tanyosphaeridium xanthiopyxides* (Wetzel emend. Morgenroth) Stover & Evitt, 1978

Thalassiphora sp.

Thalassiphora pelagica (Eisenack) Eisenack & Gocht, 1960

(Plate 6.4)

**Trigonopyxidia ginella* (Cookson & Eisenack) Downie & Sarjeant, 1965

Trithyrodinium evittii Drugg, 1967

(Plate 6.5)

Vozzhennikovia apertura (Wilson) Lentin & Williams, 1976

(Plate 6.6)

Xenascus sp. indet

(Plate 6.7)

**Xylochoarion hacknessense* Erkman & Sarjeant, 1978

7.1.2 – Acritarchs & Other Algae

Comasphaeridium cometes (Valensi) De Coninck, 1969

Cyclopsiella sp.

Cyclopsiella trematophora (Cookson & Eisenack) Lentin & Williams, 1977

Forma T

Micrhystridium sp.

(Plate 4.1)

Enigmadinium cylindfloriferum Wrenn & Hart, 1988

(Plate 2.6)

Nummus cf. *monoculatus* Morgan, 1975

indet. *Palambages*

Palambages Forma B Manum & Cookson, 1964

Palambages morulosa Wetzel, 1961

Paralecaniella indentata (Deflandre & Cookson) Cookson & Eisenack emend. Elsik

(Plate 4.6)

Veryhachium Denuff emend. Downie & Sarjeant, 1963

7.2 - Terrestrial species lists

7.2.1 – Bryophytic Spores

Foraminisporis sp. indet.

Stereisporites sp. indet.

Stereisporites antiquasporites (Wilson & Webster) Dettmann, 1963

Stereisporites regium (Drozhastchich) Drugg, 1967

7.2.2 – Lycophytic Spores

Camarozonosporites sp. indet.

(Plate 7.4)

Camarozonosporites ambigens (Fradkina) Playford, 1971

Camarozonosporites ohaiensis (Couper) Dettmann & Playford, 1968

Ceratosporites sp. indet.

Ceratosporites equalis Cookson & Dettmann, 1958

(Plate 7.5)

Retitriletes austroclavatidites (Cookson) von Döring, Krutsch, Mai & Schulz, 1963

Retitriletes eminulus (Dettmann) Srivastava, 1975

Retitriletes nodosus (Dettmann) Srivastava, 1975

(Plate 11.4)

7.2.3 – Filicean Spores

Azolla Lamarck, 1783

(Plate 7.1)

Baculatisporites sp. indet.

(Plate 7.2)

Baculatisporites comaumensis (Cookson) Potonié, 1956

Biretisporites eneaboaensis Backhouse, 1978

Biretisporites spectabilis Dettmann, 1963

Chasmatosporites sp. indet.

Clavifera triplex Bolkhovitina, 1966

Concavisporites sinuatus (Couper) Krutsch, 1959

Cyathidites sp. indet.

(Plate 7.6)

Deltoidospora sp. have also been used. Where *Deltoidospora* sp. is used, *Cyathidites* sp. is intended.

Dictyophyllidites sp. indet.

Dictyophyllidites concavus Harris, 1965

Dictyophyllidites harrisi Couper, 1958

Gleicheniidites circinidites (Cookson) Brenner, 1963

(Plate 8.2)

Gleicheniidites concavus

Ischyosporites sp. indet.

Ischysporites gremius Stover, 1973

(Plate 8.4)

Laevigatosporites sp. indet.

Laevigatosporites granulosus Couper, 1953

Laevigatosporites major (Cookson) Krutsch, 1959

Laevigatosporites ovatus Wilson & Webster, 1946

(Plate 8.5)

Osmundacidites sp.

Osmundacidites wellmannii Couper, 1953

Polypodiisporites sp.

Polypodiisporites speciosus (Harris) Kahn & Martin, 1971

(Plate 11.1)

Triletes verrucatus Couper, 1953

(Plate 11.7)

7.2.4 – Gymnospermous Pollen

Araucariacites australis Cookson, 1947

Cycadopites sp.

(Plate 7.7)

Cycadopites minor Couper, 1953

Dacrydiumites sp. indet.

Dacrydiumites bidwillii Cookson, 1940

Dacrydiumites florinii Cookson & Pike, 1953

Dacrydiumites franklinii Cookson, 1953

Inaperturopollenites sp. indet.

Inaperturopollenites australis Couper, 1953

Microcachydites sp. indet.

Microcachydites antarcticus Cookson ex Couper, 1953

(Plate 8.8)

Phyllocladus sp. indet.

Phyllocladidites mawsonii Cookson ex Couper, 1953

(Plate 10.7)

Podocarpidites spp

Podocarpus dacrydioides Rich, 1953

(Plate 10.8)

7.2.5 – Angiospermous Pollen

Aequitriradites sp. indet.

Aequitriradites verrucosus (Cookson & Dettmann) Cookson & Dettmann, 1961

Arecipites sp.

Beaupreacidites sp.

(Plate 7.3)

Beaupreacidites verrucosus Cookson, 1950

Clavamonocolpites polygonalis Askin, 1994

Gemmamonocolpites sp. indet

Gemmamonocolpites pilulus Askin, 1994

(Plate 8.1)

Haloragacidites sp. indet.

Haloragacidites harrisii (Couper) Harris, 1971

(Plate 8.3)

Liliacidites sp.

Liliacidites sp. 1 Askin, 1994

(Plate 8.6)

Liliacidites aviemorensis McIntyre, 1968

Liliacidites kaitangataensis Couper, 1953

(Plate 8.7)

Liliacidites variegatus Couper, 1953

Monosulcites sp. indet.

Nothofagidites sp. indet

Nothofagidites acromegacanthus Menéndez & Caccavari, 1975

Nothofagidites americanus del Zamaloa, 1992

(Plate 9.1)

Nothofagidites astrus (Couper) Hekel, 1972

(Plate 9.2)

Nothofagidites brachyspinulosus (Cookson) Harris, 1965

(Plate 9.3)

Nothofagidites cinctus (Cookson) Fasola, 1969

Nothofagidites cranwelliae (Couper) Mildenhall & Pocknall, 1989

Nothofagidites deminutus (Cookson) Stover & Evans, 1973

Nothofagidites dorotensis Romero, 1973

(Plate 9.4)

Nothofagidites emarcidus (Cookson) Harris, 1965

(Plate 9.5)

Nothofagidites falcatus (Cookson) Hekel, 1972

Nothofagidites flemingii (Couper) Potonié, 1960

Nothofagidites fuegensis Menéndez & Caccavari, 1975

(Plate 9.6)

Nothofagidites goniatus (Cookson) Stover & Evans, 1973

(Plate 9.7)

Nothofagidites heterus (Cookson) Stover & Evans, 1973

Nothofagidites incrassatus (Cookson) Dettmann, 1990

(Plate 9.8)

Nothofagidites lachlaniae (Couper) Pocknall & Mildenhall, 1984

(Plate 10.1)

Nothofagidites matauraensis (Couper) Hekel, 1972

(Plate 10.2)

Nothofagidites nanus Romero, 1977

Nothofagidites rocaensis Romero, 1973

(Plate 10.3)

Nothofagidites saraensis Menéndez & Caccavari, 1975

(Plate 10.4)

Nothofagidites senectus Dettmann & Playford, 1968

Nothofagidites spinosus (Couper) Mildenhall & Pocknall, 1989

(Plate 10.5)

Nothofagidites suggatei (Couper) Hekel, 1972

Nothofagidites visserensis Romero, 1973

Peninsulapollis sp.

Peninsulapollis gillii Cookson emend. Dettmann & Jarzen, 1988

(Plate 10.6)

Peninsulapollis truswelliae Dettmann & Jarzen, 1988

Peromonolites sp. indet.

Peromonolites bowenii Couper, 1953

Plicatipollenites malabarensis (Potonié & Sah) Foster, 1976

Propylipollis sp.

(Plate 11.2)

Proteacidites sp.

(Plate 11.3)

Sparganiaceaepollenites irregularis Kemp, 1977

Tricolpites sp

(Plate 11.5)

Tricolporopollenites sp.

(Plate 11.6)

Triporopollenites sp.

(Plate 11.8)

These pages
have been p/copied

Plate 1

All magnifications approximately x500 unless otherwise stated

1. Alisocysta circumtabulata

Sample DJ 904.45 A (P67/4). López de Bertodano Formation

2. Alisocysta reticulata

Sample DJ 904.55 A (P62.3). López de Bertodano Formation

3. Alisocysta rugolirata

Sample DJ 904.43 A (N63/4-N64.3). López de Bertodano Formation

4. Arachnodinium antarcticum

Sample DJ 802.27 A (L61/3-0). La Meseta Formation

magnification x450

5. Bourkidinium cylinderatum

Sample DJ 904.42 B (H43/0). López de Bertodano Formation

magnification x750

6. *Ceratiopsis striata*

Sample DJ 904.57 A (J64/2). López de Bertodano Formation

7. Cerodinium speciosa

Sample DJ 801.1 B (T41/1). Sobral Formation

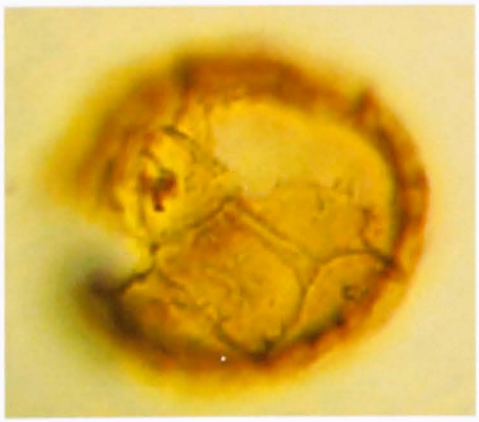
magnification x450

8. *Cordosphaeridium minimum*

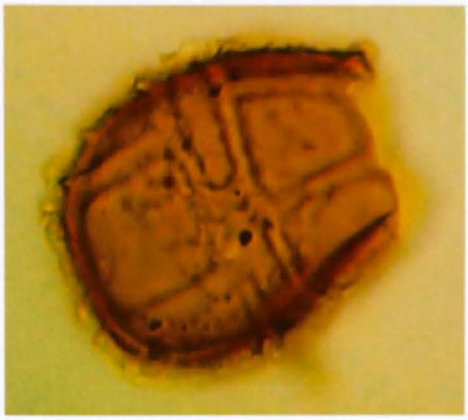
Sample DJ 904.57 A (P65/2). López de Bertodano

magnification x750

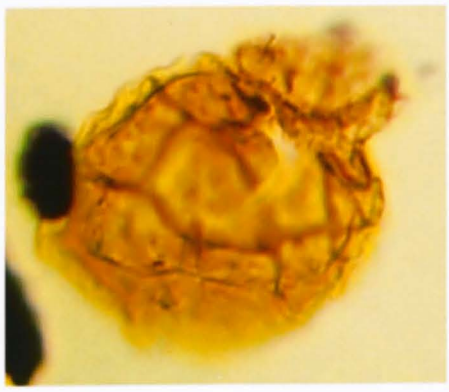
Plate 1



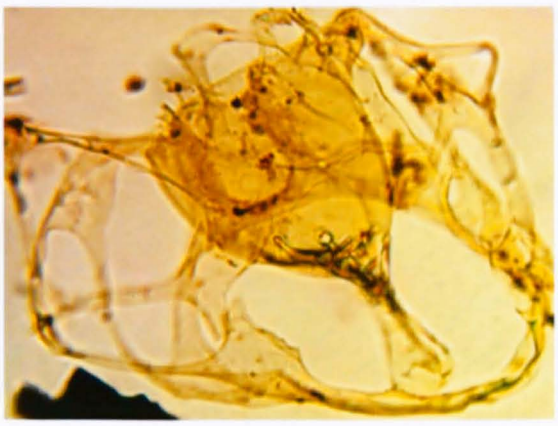
1



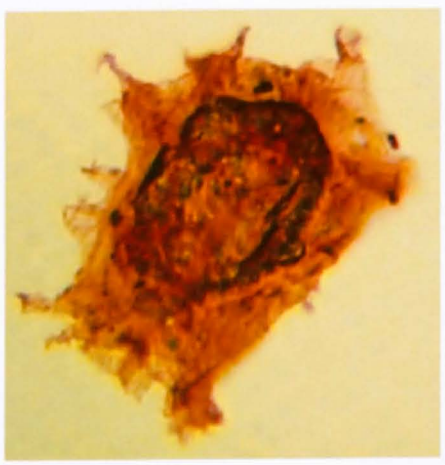
2



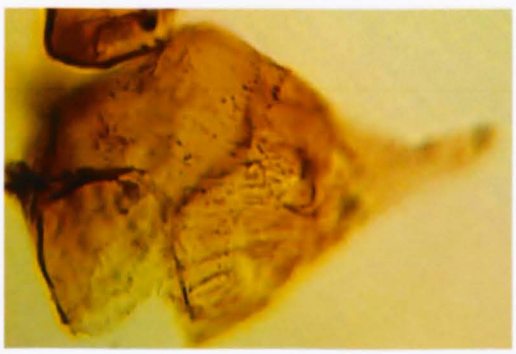
3



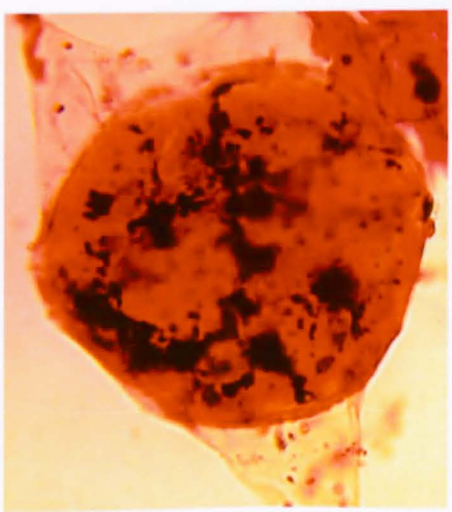
4



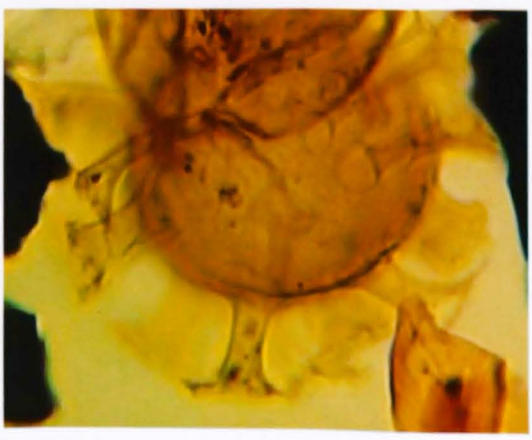
5



6



7



8

Plate 2

All magnifications approximately x500 unless otherwise stated

1. Cometodinium whitei

Sample DJ 802.25 A (M66/1). La Meseta Formation

2. Cribroperidineum edwardsii

Sample DJ 802.20A B (K39/2). La Meseta Formation

magnification x450

3. Deflandrea antarctica

Sample DJ 802.34 B (Q47/0-1). La Meseta Formation

magnification x450

4. Deflandrea medcalfii

Sample DJ 904.53 B (M52/0). López de Bertodano Formation

magnification x450

5. Diconodinium multispinulum

Sample DJ 904.42 A (G67/3). López de Bertodano Formation

6. Enigmadinium cylindfloriferum

Sample DJ 802.34 B (P46/0-P46/1). La Meseta Formation

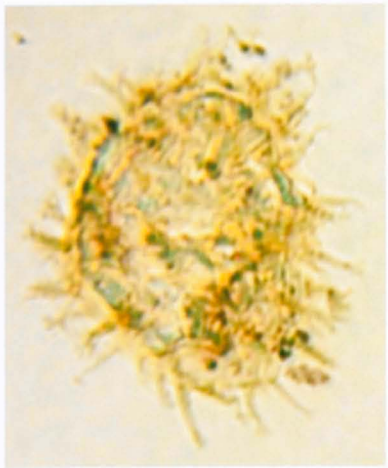
7. Enneadocysta partridgei

Sample DJ 802.34 B (N51/0). La Meseta Formation

8. Enneadocysta partridgei sub-species A

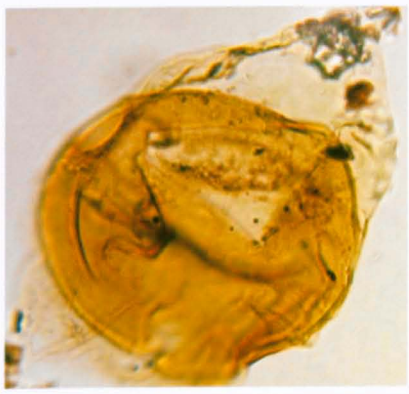
Sample DJ 802.32 C (N31/0). La Meseta Formation

Plate 2

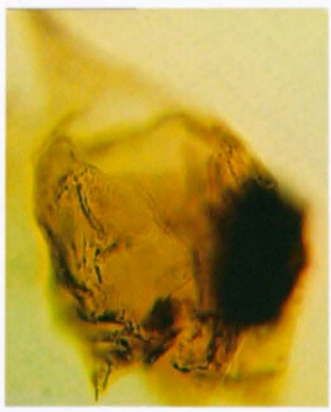


1

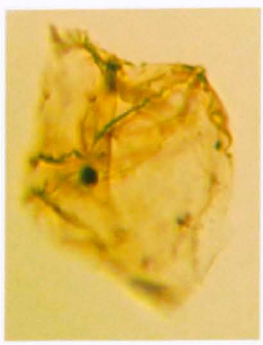
2



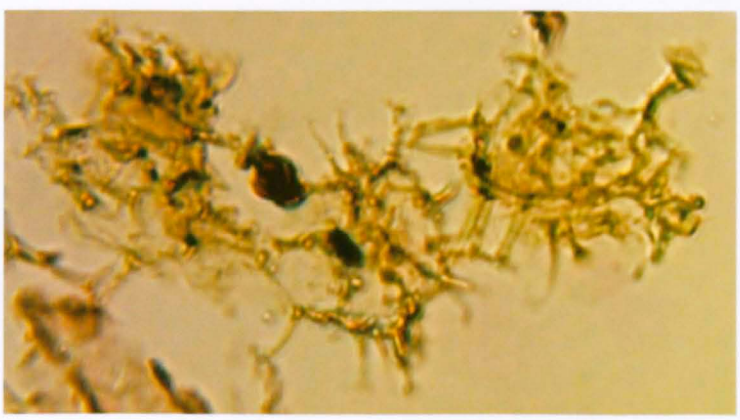
3



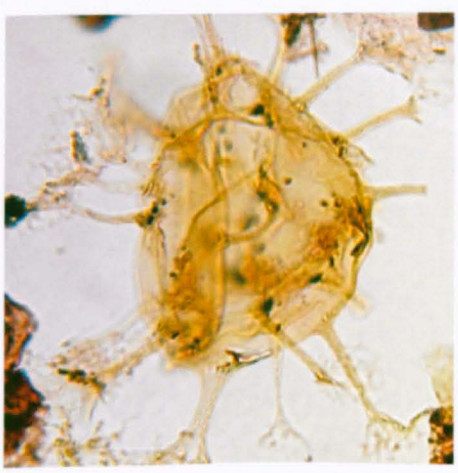
4



5



6



7

8

Plate 3

All magnifications approximately x500 unless otherwise stated

1. Exosphaeridiidium bifidum

Sample DJ 904.43 A (R62/0). López de Bertodano Formation

2. Hystrichosphaeridium salpingophorum

Sample DJ 904.40 B (N43/0). López de Bertodano Formation

3. Hystrichosphaeridium truswelliae

Sample DJ 802.27 A (Q65/0-1). La Meseta Formation

4. Hystrichosphaeridium tubiferum

Sample DJ 904.57 B (Q42/2). López de Bertodano Formation

5. Impagidinium maculatum

Sample DJ 802.26 B (N64/0). La Meseta Formation

6. *Impagidinium victorianum*

Sample DJ 802.28 B (P46/2). La Meseta Formation

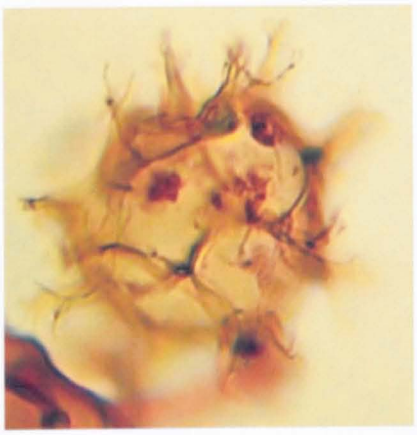
7. Impletosphaeridium sp.

Sample DJ 802.16 A (N65/0). La Meseta Formation

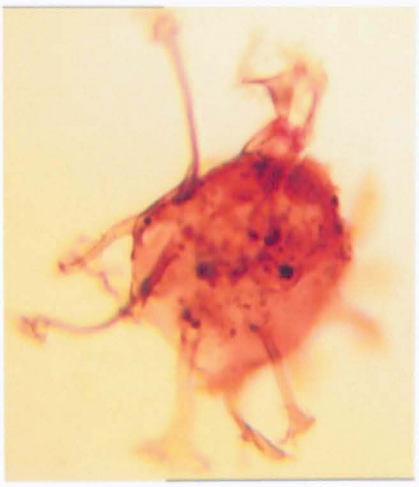
8. *Lejeunecysta fallax*

Sample DJ 802.24 C (K37/0). La Meseta Formation

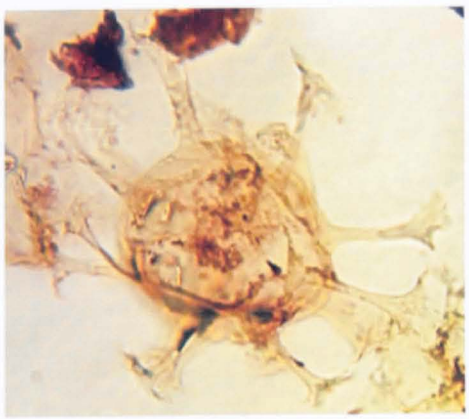
Plate 3



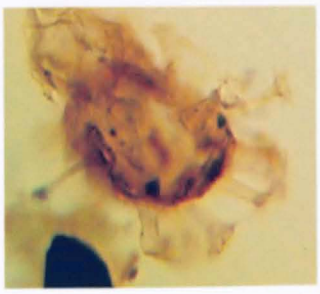
1



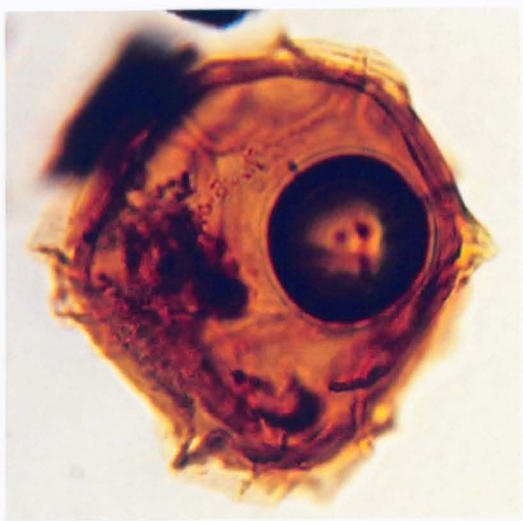
2



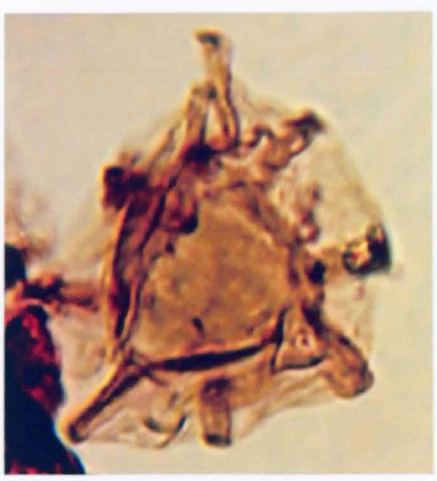
3



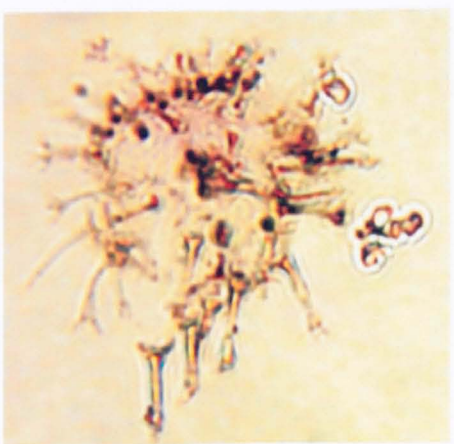
4



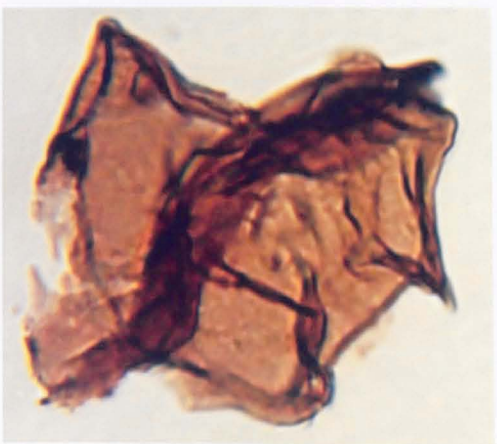
5



6



7



8

Plate 4

All magnifications approximately x500 unless otherwise stated

1. Micrhystridium sp.

Sample DJ 802.14 B (M44/2). La Meseta Formation

magnification x750

2. Microdinium sp.

Sample DJ 904.53 B (M50/0-3). López de Bertodano Formation

magnification x750

3. Octodinium askiniae

Sample DJ 802.24 A (K68/1-0). La Meseta Formation

4. Palaeocystodinium golzowense

Sample DJ 904.52 B (o44/2-o45/1). López de Bertodano Formation

5. Palaeoperidinium pyrophorum

Sample DJ 801.1 A (P62/0). Sobral Formation

6. Paralecaniella indentata

Sample DJ 802.14 B (K46/3). La Meseta Formation

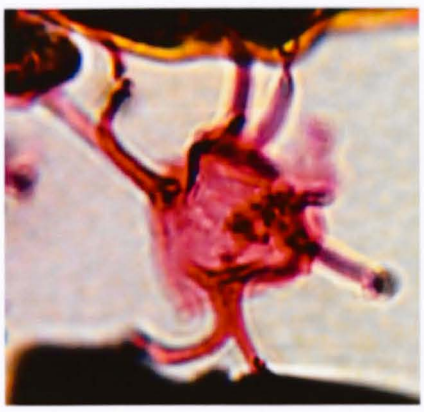
7. Pyxidiniopsis delicata

Sample DJ 802.27 A (O63/2-N63/4). La Meseta Formation

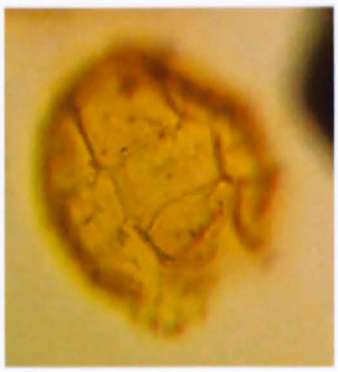
8. Pyxidiniopsis crassimurata

Sample DJ 904.43 A (Q59/0). López de Bertodano Formation

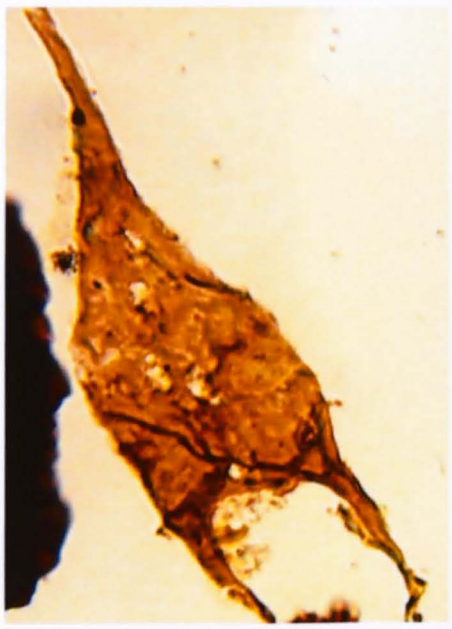
Plate 4



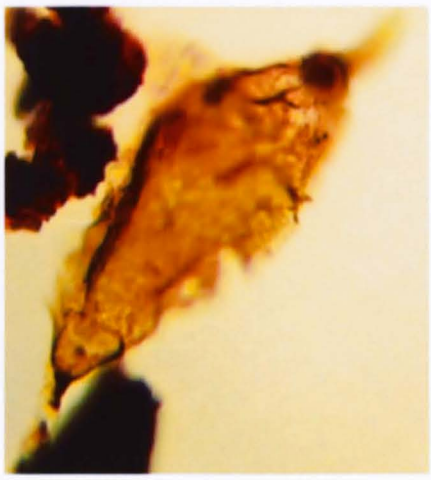
1



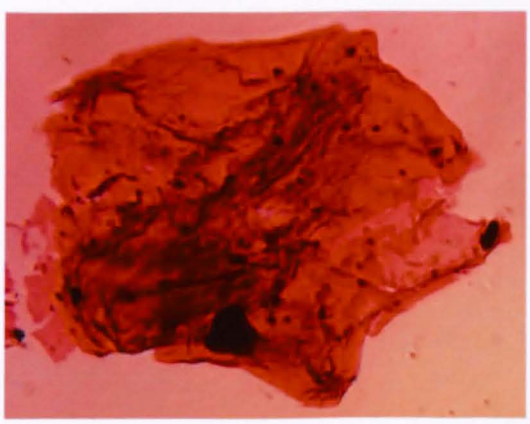
2



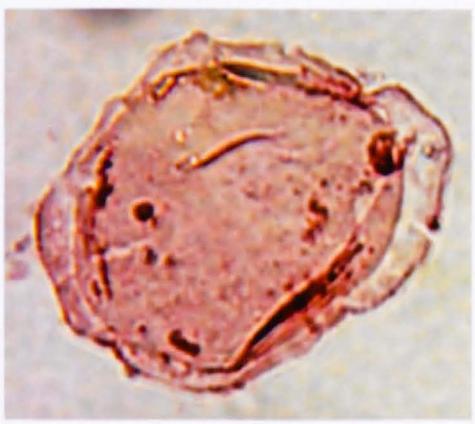
3



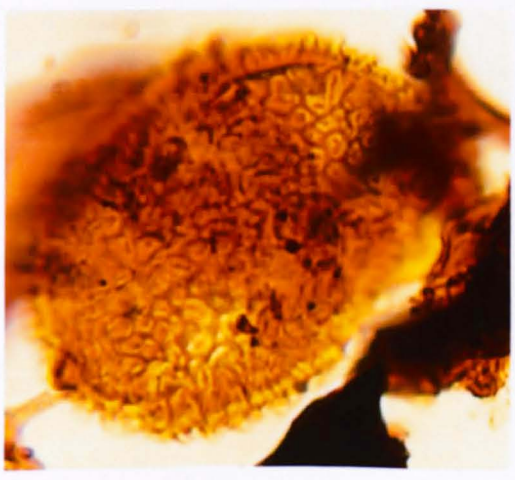
4



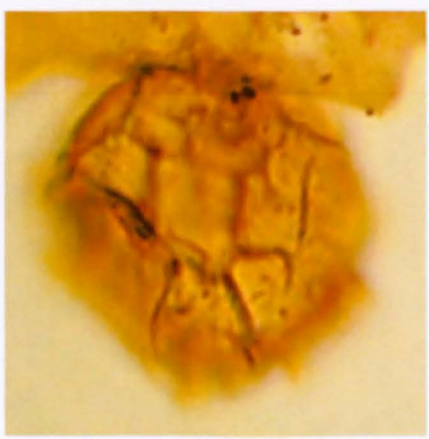
5



6



7



8

Plate 5

All magnifications approximately x500 unless otherwise stated

1. *Pseodoceratium securigera*

Sample DJ 802.29 C (M32/0). La Meseta Formation

2. *Selenopemphix nephroides*

Sample DJ 802.25 A (L64/1). La Meseta Formation

3. *Sengalinium dilwynense*

Sample DJ 904.64 B (O44/0-1). López de Bertodano Formation

4. *Spinidinium colemanii*

Sample DJ 803.7 B (N52/4). La Meseta Formation

5. *Spinidinium essoi*

Sample DJ 802.27 C (N66/3). La Meseta Formation

6. *Spindinium luciae*

Sample DJ 802.10 A (O63/3). La Meseta Formation

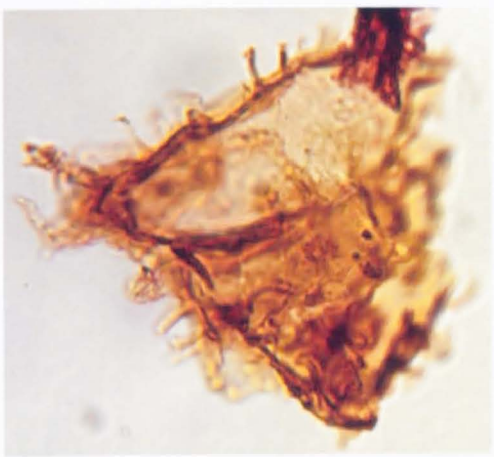
7. *Spinidinium macmurdoense*

Sample DJ 803.6 B (L52/1-3). La Meseta Formation

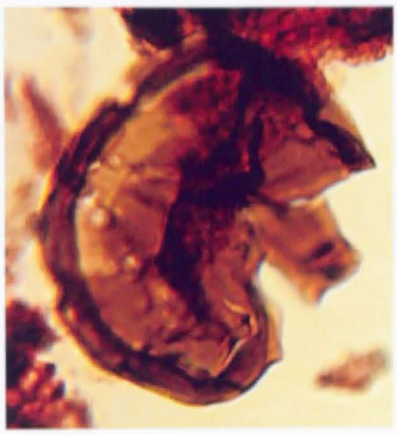
8. *Spinidinium pilatum*

Sample DJ 904.41 B (R46/2-R47/1). López de Bertodano Formation

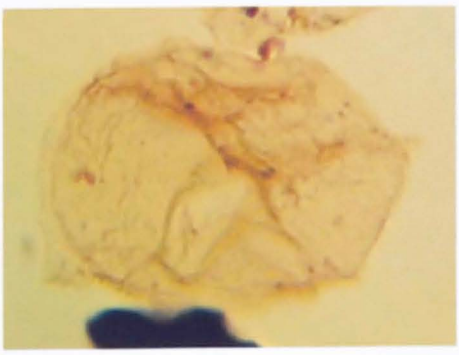
Plate 5



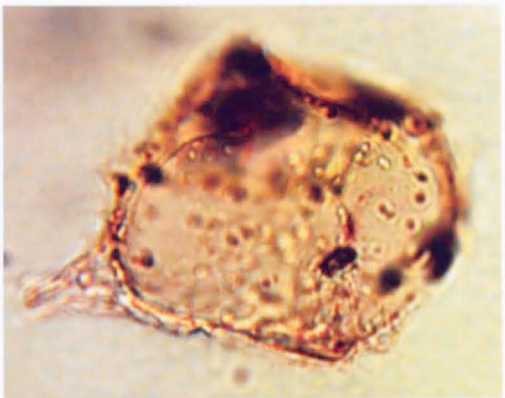
1



2

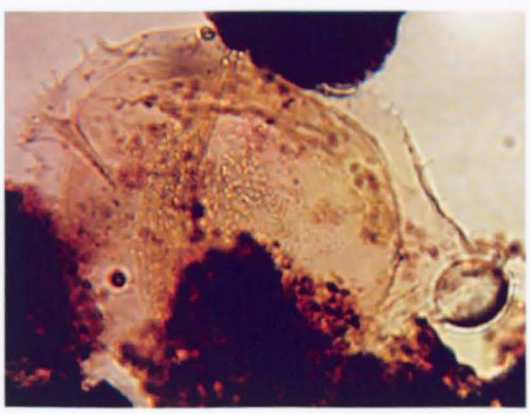


3

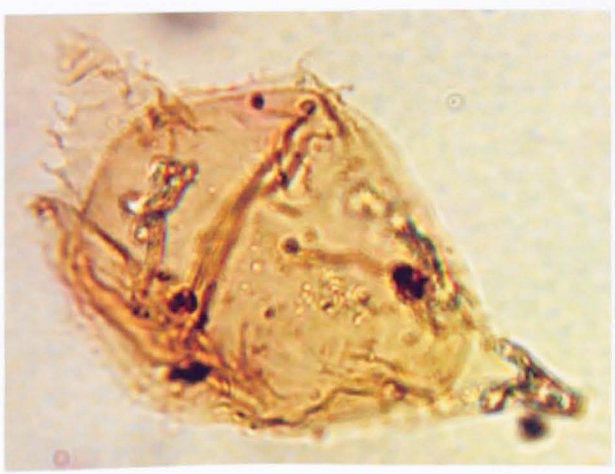


4

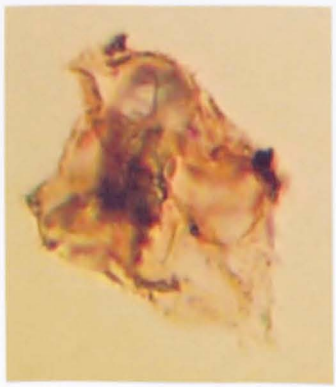
5



6



7



8

Plate 6

All magnifications approximately x500 unless otherwise stated

1. Spiniferites microceras

Sample DJ 802.12 C (H31/0). La Meseta Formation

2. Spiniferites pseudofurcatus

Sample DJ 904.40 A (M69/0-4). López de Bertodano Formation

3. Subtilisphaera sp.

Sample DJ 904.39 A (O61/3-P61/1). López de Bertodano Formation

4. Thalassiphora pelagica

Sample DJ 802.27 A (O63/0). La Meseta Formation

magnification x450

5. Trithyrodinium evittii

Sample DJ 904.52 B (J48/1). López de Bertodano Formation

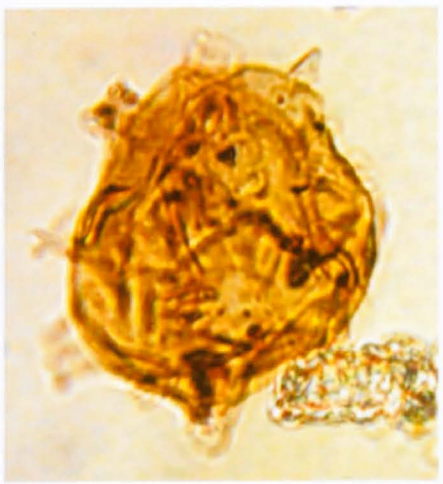
6. Vozzhennikovia apertura

Sample DJ 802.32 A (P61/3-4). La Meseta Formation

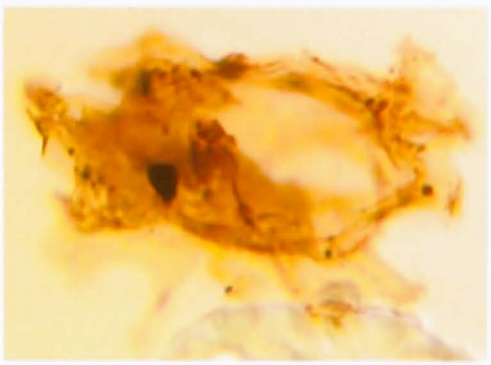
7. Xenascus sp.

Sample DJ 904.42 B (J50/1). López de Bertodano Formation

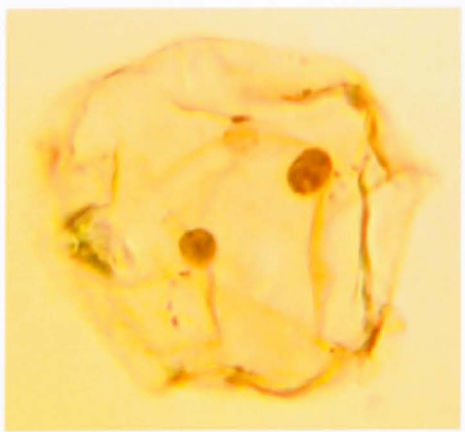
Plate 6



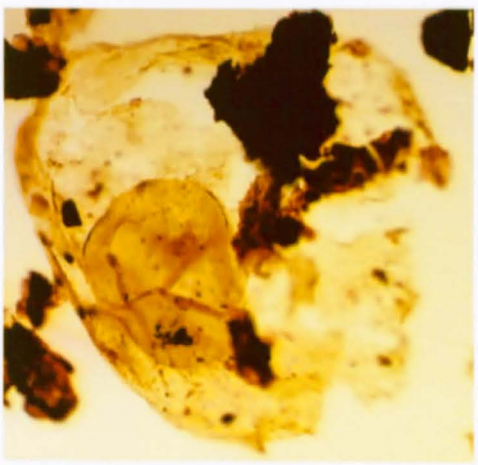
1



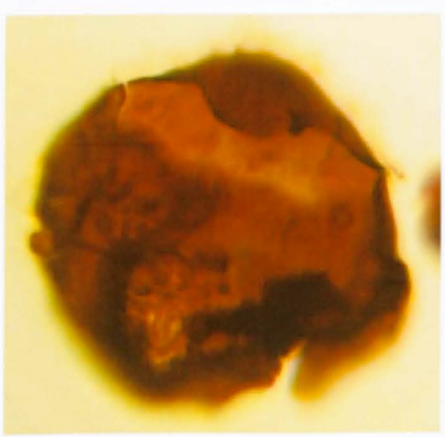
2



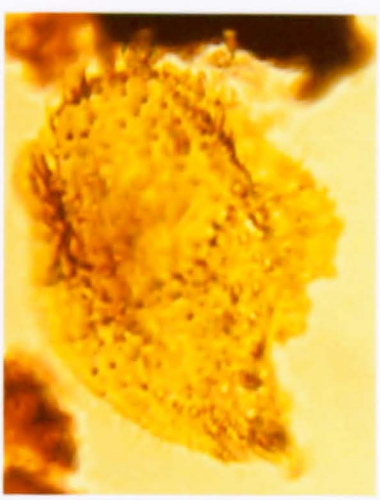
3



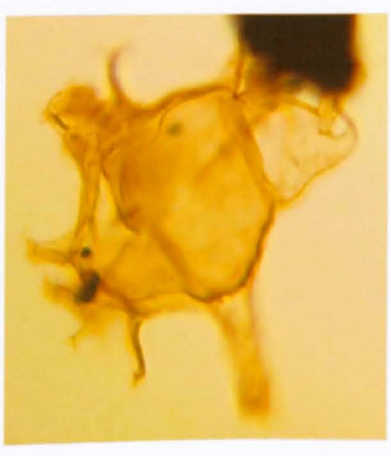
4



5



6



7

Plate 7

All magnifications approximately x750 unless otherwise stated

1. Azolla sp.

Sample DJ 904.44 A (L64/0). López de Bertodano Formation

2. Baculatisporites sp

Sample DJ 802.26 B (K53/2). López de Bertodano Formation

3. Beaupreacidites sp

Sample DJ 904.63 A (K72/3-L72/3). López de Bertodano Formation

4. Camarozonosporites sp.

Sample DJ 802.23 B (L51/4-M51/2). La Meseta Formation

5. Ceratosporites equalis

Sample DJ 802.12 A (J61/4-J62/3). La Meseta

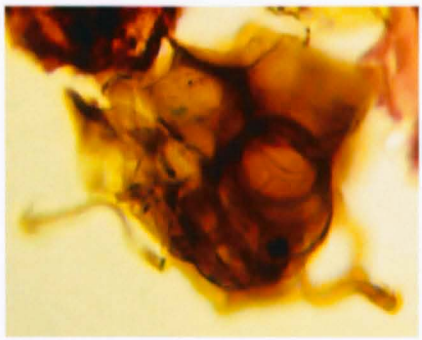
6. Cyathidites spp.

Sample DJ 802.24 A (M66/1). La Meseta Formation

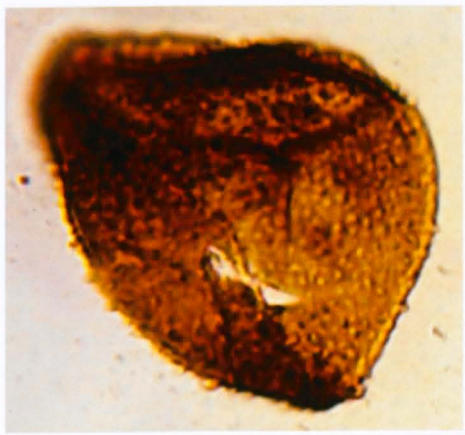
7. Cycadopites sp.

Sample DJ 904.58 A (O62/4). López de Bertodano Formation

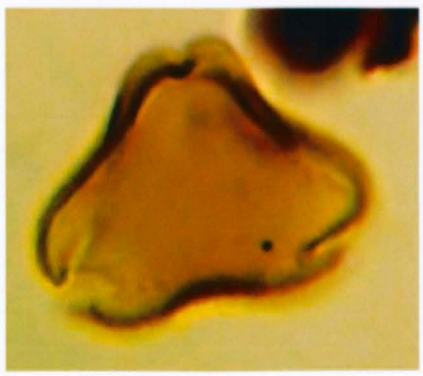
Plate 7



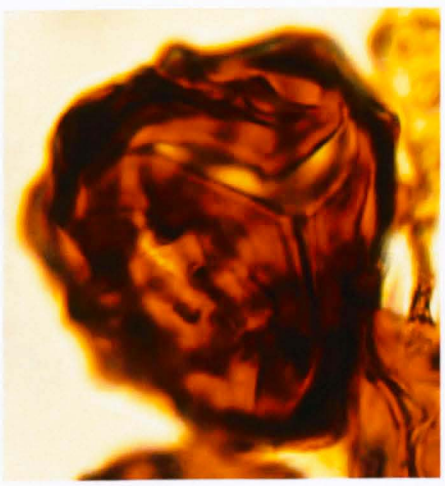
1



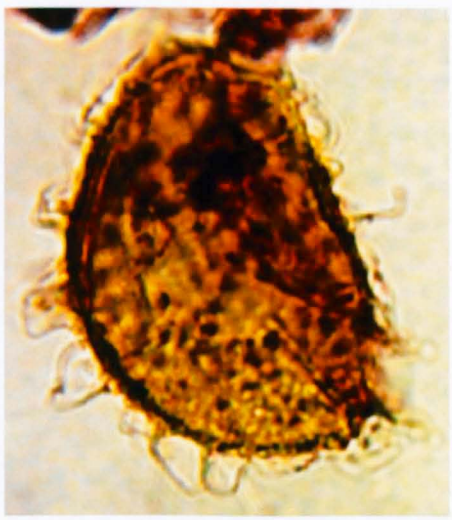
2



3

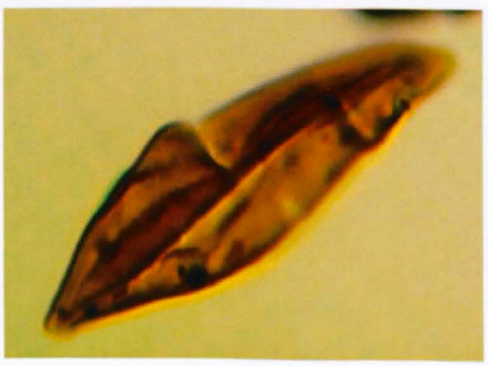


4



5

6



7

Plate 8

All magnifications x750 unless otherwise stated

1. *Gemmamonocolpites pilulus*

Sample DJ 904.40 A. López de Bertodano Formation

2. *Gleicheniidites circinidites*

Sample DJ 802.23 A (N71/1). La Meseta Formation

3. *Haloragacidites harrisii*

Sample DJ 904.44 B (O43/0). López de Bertodano Formation

4. *Ischyosporites gremius*

Sample DJ 802.24 B (O50). La Meseta Formation

5. *Laevigatosporites ovatus*

Sample DJ 904.54 A (M60/0). López de Bertodano Formation

6. *Liliacidites* sp. 1 (Askin, 1988)

Sample DJ 904.43 A (N62/1). López de Bertodano Formation

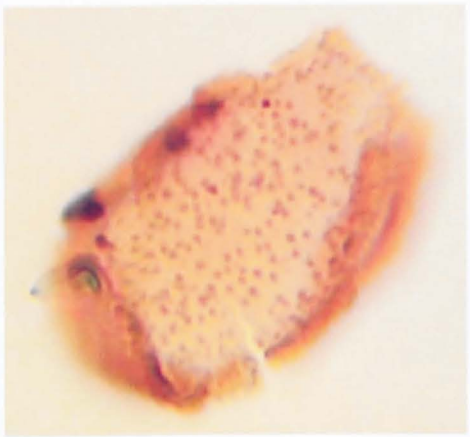
7. *Liliacidites kaitangatensis*

Sample DJ 904.45 A (P63/4). López de Bertodano Formation

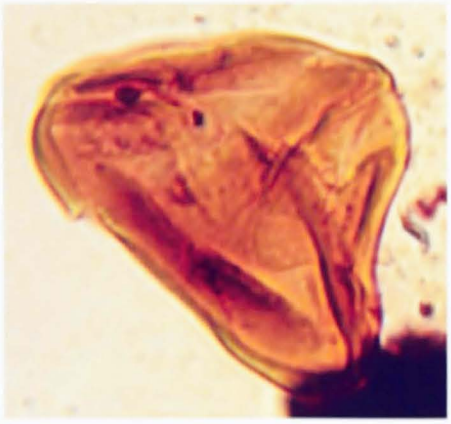
8. *Microcachrydites antarcticus*

Sample DJ 904.41 B (H46/4). López de Bertodano Formation

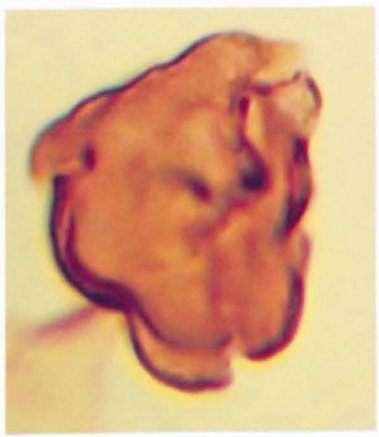
Plate 8



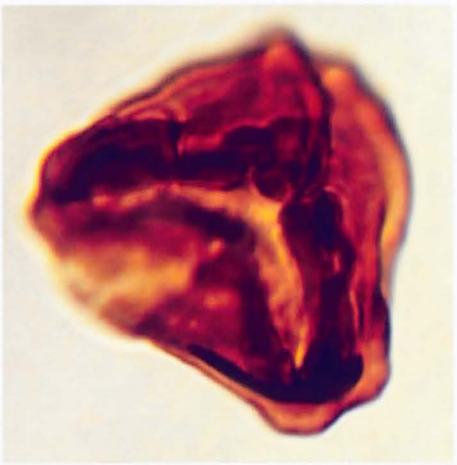
1



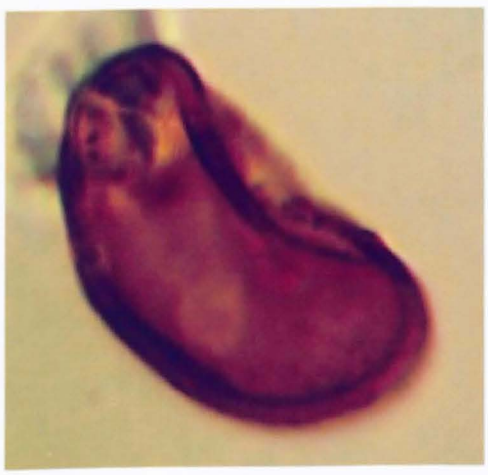
2



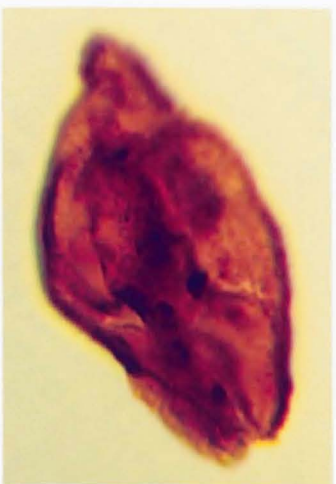
3



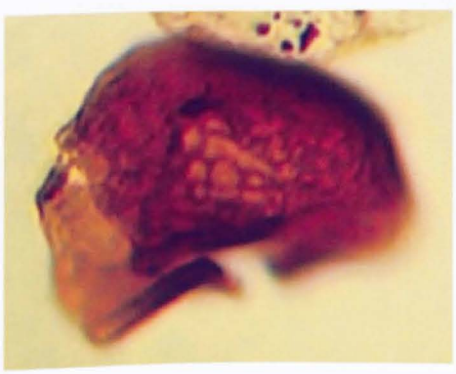
4



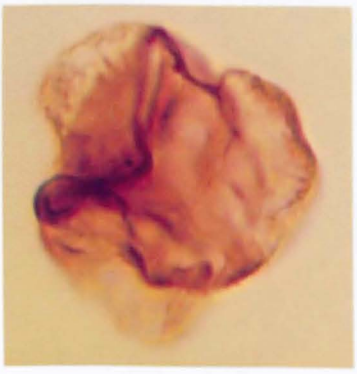
5



6



7



8

Plate 9

All magnifications x750 unless otherwise stated

1. *Nothofagidites americanus*

Sample DJ 802.20 A (P63/0). La Meseta Formation

2. *Nothofagidites astrus*

Sample DJ 802.14 B (P51/0-3). La Meseta Formation

3. *Nothofagidites brachyspinulosus*

Sample DJ 802.19 C (R33/4-0). La Meseta Formation

4. *Nothofagidites dorotensis*

Sample DJ 802.19 C (Q28/2). La Meseta Formation

5. *Nothofagidites emarcidus*

Sample DJ 802.25 A (M61/0-2). La Meseta Formation

6. *Nothofagidites fuegensis*

Sample DJ 802.19 B (P47/0-2). La Meseta Formation

7. *Nothofagidites goniatus*

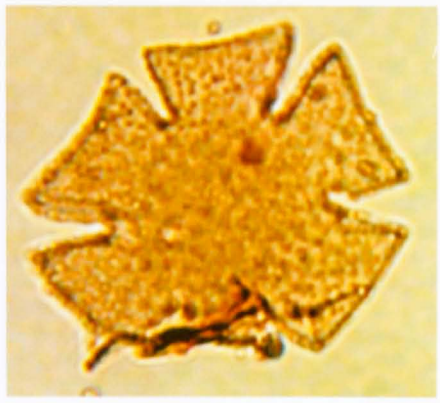
Sample DJ 802.15 C (K30/0). La Meseta Formation

8. *Nothofagidites incrassatus*

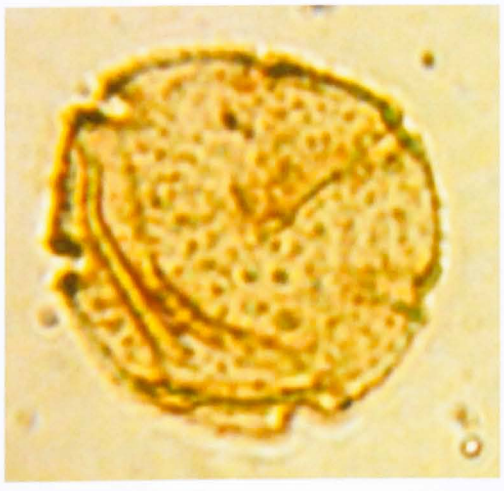
Sample DJ 802.14 A (L66/4-M66/2). La Meseta Formation

Plate 9

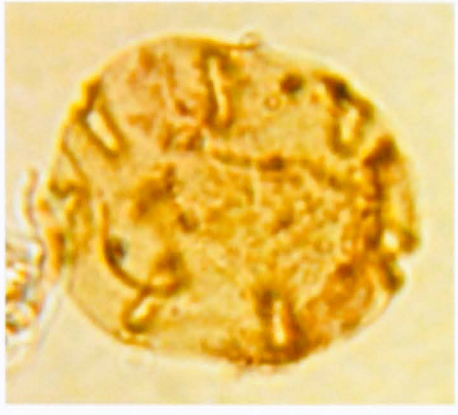
1



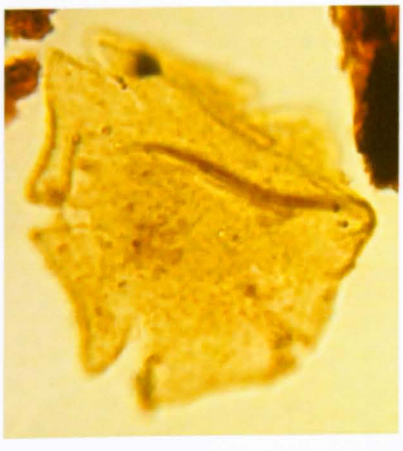
2



3

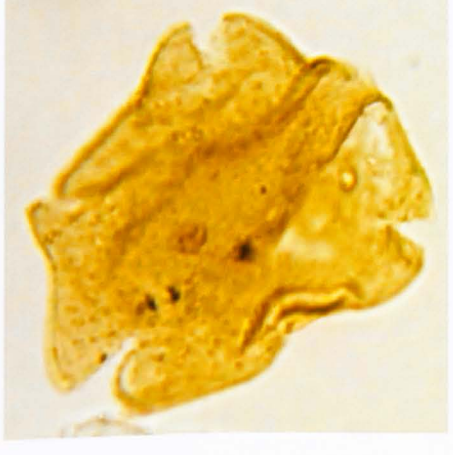


4

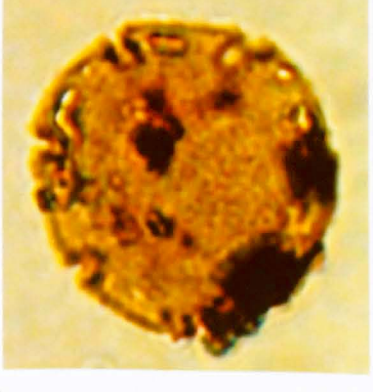


5

6



7



8

Plate 10

All magnifications x750 unless otherwise stated

1. Nothofagidites lachlaniae

Sample DJ 802.26 B (R46/0-2). La Meseta Formation

2. Nothofagidites mataurаensis

Sample DJ 802.19 C (R27/2). La Meseta Formation

3. Nothofagidites rocaensis

Sample DJ 802.16 A (P66/0). La Meseta Formation

4. Nothofagidites saraensis

Sample DJ 802.16 B (O46/0). La Meseta Formation

5. Nothofagidites spinosus

Sample DJ 802.16 B (N50/0). La Meseta Formation

6. Peninsulapollis gillii

Sample DJ 801.1 B (U40/2). La Meseta Formation

7. Phyllocladidites mawsonii

Sample DJ 802.19 A (M68/3). La Meseta Formation

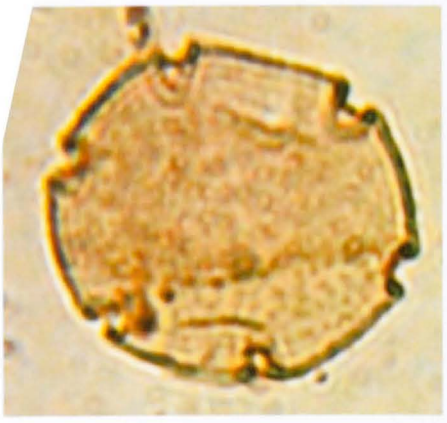
magnification x500

8. Podocarpus dacrydioides

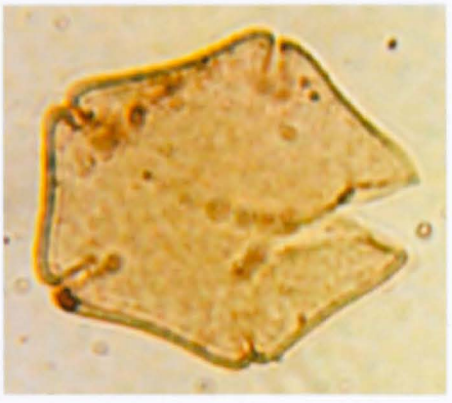
Sample DJ 802.23 C (O31/0). La Meseta Formation

magnifiation x500

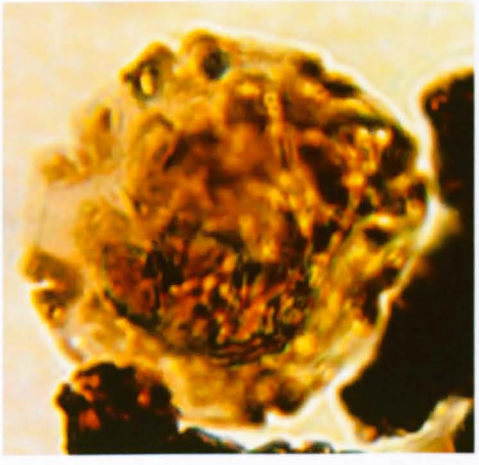
Plate 10



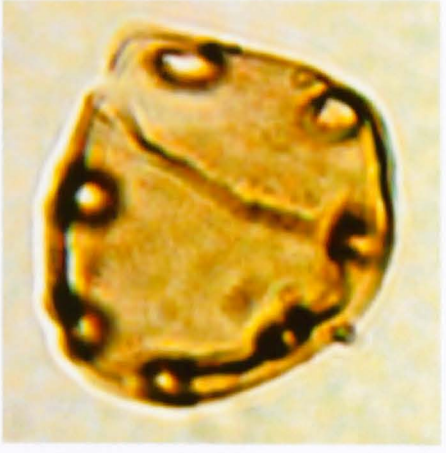
1



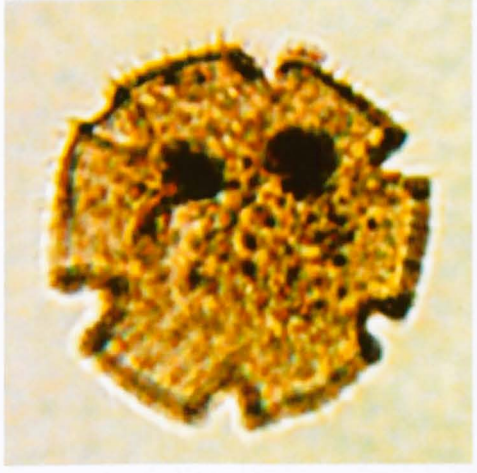
2



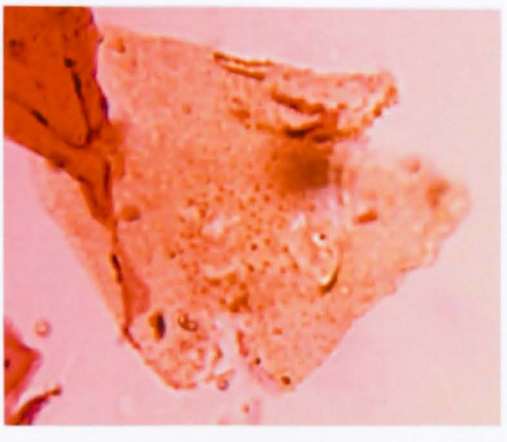
3



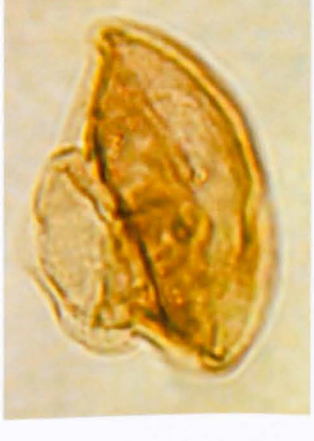
4



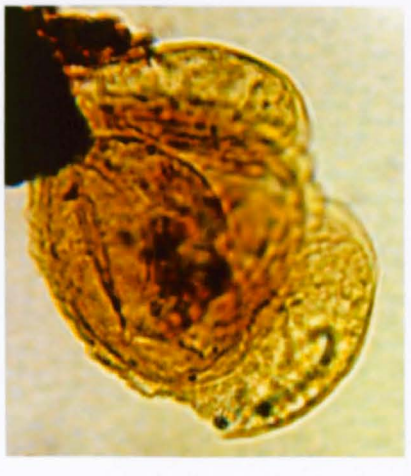
5



6



7



8

Plate 11

All magnifications x750 unless otherwise stated

1. Polypodiisporites speciosus

Sample DJ 803.6 B (L52/2-4-L53/1-3). La Meseta Formation

2. Propylipollis sp.

Sample DJ 904.45 A (L64/4). López de Bertodano Formation

3. Proteacidites sp.

Sample DJ 807.6 A (L61/2). La Meseta Formation

4. Retitriletes nodosus

Sample DJ 904.41 B (K44/3-4). López de Bertodano Formation

5. Tricolpites sp.

Sample DJ 904.40 A (O65/4-P65/2). López de Bertodano Formation

6. Tricolporopollenites sp.

Sample DJ 904.43 A (P60/3). López de Bertodano Formation

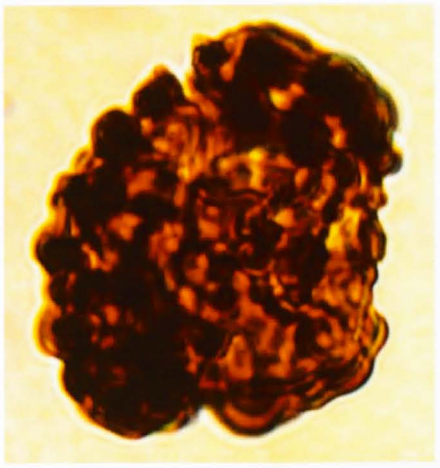
7. Triletes verrucatus

Sample DJ 802.25 A (N68/3-0). La Meseta Formation

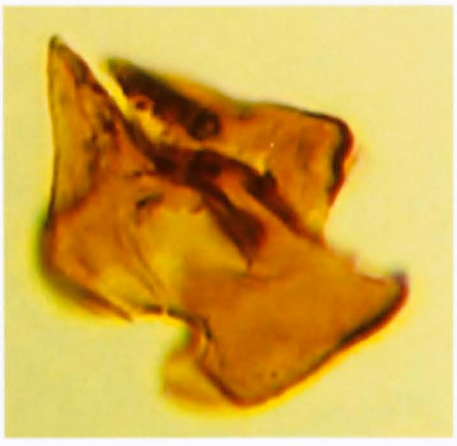
8. Triporopollenites sp.

Sample DJ 904.41 A (J61/0-4). López de Bertodano Formation

Plate 11

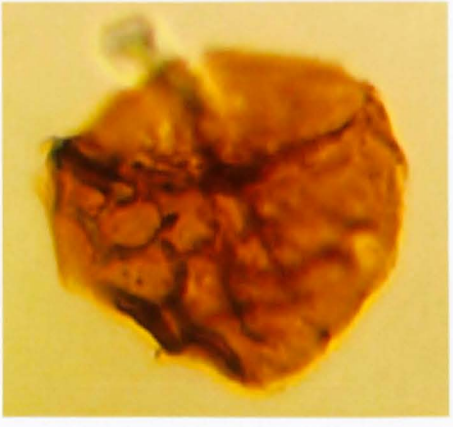


1

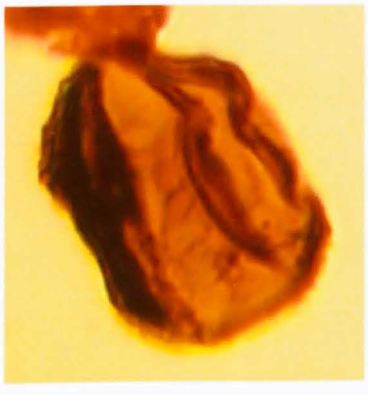


2

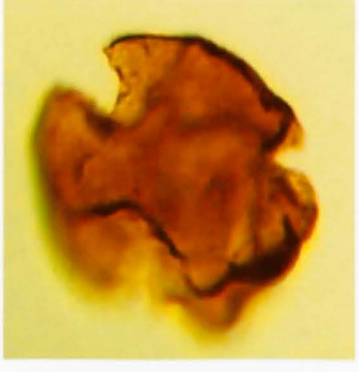
3



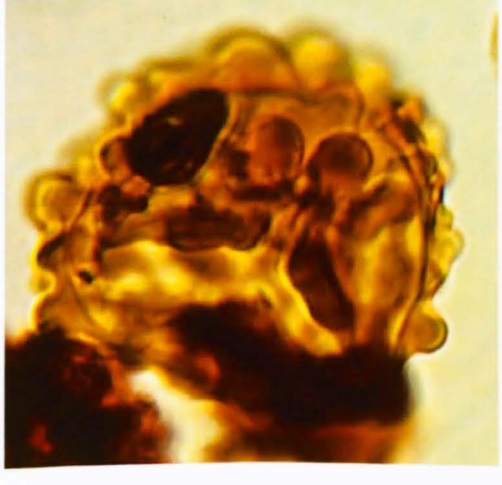
4



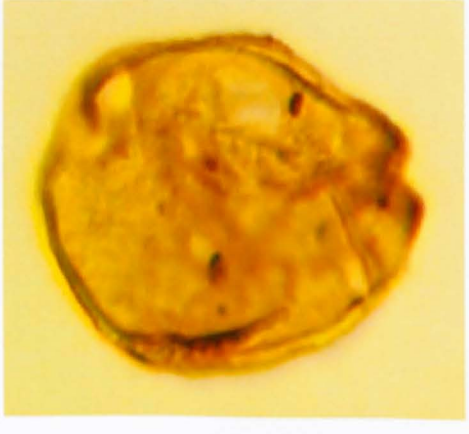
5



6



7



8

REFERENCES

Aguirre-Uretta, M.B., Buatois, L.A., Chernoglasov, G.Ch.B. & Medina, F.A., 1990. First Polychelidae (Crustacea, Palinura) from the Jurassic of Antarctica. *Antarctic Science*, **2**, 157-162.

Askin, R.A., 1988a. Campanian to Paleocene palynological succession of Seymour and adjacent islands, northeastern Antarctic Peninsula. *In*: Feldmann, R.M. & Woodburne, M.O., (eds). *Geology and Paleontology of Seymour Island, Antarctic Peninsula.* Geological Society of America Memoir, **169**, 131-153.

Askin, R.A., 1988b. The palynological record across the Cretaceous/Tertiary transition on Seymour Island, Antarctica. *In*: Feldmann, R.M. & Woodburne, M.O.,(eds). *Geology and Paleontology of Seymour Island, Antarctic Peninsula.* Geological Society of America Memoir, **169**, 155-162

Askin, R.A., 1989. Endemism and heterochroneity in the Late Cretaceous (Campanian) to Paleocene palynofloras of Seymour Island, Antarctica: implications for origins, dispersal and palaeoclimates of southern floras. *Geological Society Special Publication*, **47**, 107-119.

Askin, R.A., 1990. Campanian to Paleocene spore and pollen assemblages of Seymour Island, Antartica. *Review of Palaeobotany and Palynology*, **65**, 105-113.

Askin, R.A., Elliot, D.H., Stilwell, J.D. & Zinsmeister W.J., 1991. Stratigraphy and Paleontology of Campanian and Eocene sediments, Cockburn Island, Antarctic Peninsula. *Journal of South American Earth Sciences*, **4**, 99-117.

Barker, P.F., 1982. The Cenozoic subduction history of the Pacific margin of the Antarctic Peninsula: ridge crest-trench interactions. *Journal of the Geological Society, London*, **139**, 787-801.

Barker, P.E. & Burrell, J., 1977. The opening of the Drake Passage. *Marine Geology*, **25**, 15-34.

Barreda, V., Palamarczuk, S. & Medina, F., 1999. Palinología de la Formación Hidden Lake (Coniaciano-Santoniano), Isla James Ross, Antártida. *Revista Española de Micropaleontología*, **31 (1)**, 53-72.

Berggren, W.A. & Hollister, C.D., 1977. Plate tectonics and paleocirculation-commotion in the ocean. *Tectonophysics*, **11**, 11-48.

Brinkhuis, H., Bujak, J.P. Smit, J., Versteegh, G.J.M & Visscher, H., 1998. Dinoflagellate-based temperature reconstructions across the Cretaceous-Tertiary boundary. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **141**, 67-83.

Bujak, J.P., 1984. Cenozoic dinoflagellate cysts and acritarchs from the Bering Sea and northern North Pacific, D.S.D.P. Leg 19. *Micropaleontology*, **30**, 180-212.

Cape Roberts Science Team, 1998. Initial Report on CRP-1, Cape Roberts Project, Antarctica. *Terra Antartica*, **5**, 1-187.

Cape Roberts Science Team, 2000. Studies from the Cape Roberts Project, Ross Sea, Antarctica - Initial Report on CRP-3. *Terra Antartica*, **7**, 1-209.

Chaloner, W.G. & Muir, M., 1968. Spores and floras. *In*: Murchison, D.G. & Westall, T.S. (eds). *Coal and coal-bearing strata*. Oliver & Boyd, Edinburgh, 127-146.

Christophel, D.C., 1989. Evolution of the Australian flora through the Tertiary of South-eastern Australia. *Plant Systematics and Evolution*, **162**, 63-78.

Cookson, I. C., 1953. The identification of the sporomorph *Phyllocladidites* with *Dacrydium* and its distribution in southern Tertiary deposits. *Australian Journal of Botany*, **1**, 64-70.

Cocozza, C.D. & Clarke, C.M., 1992. Eocene microplankton from the La Meseta Formation, northern Seymour Island. *Antarctic Science*,**4**, 355-362.

Cox, K.G., 1992. Karoo igneous activity, and the early stages of the break-up of Gondwanaland. *In* Storey, B.C., Alabaster, T. & Pankhurst, R. (eds). *Magmatism and the Causes of Continental Break-up*. Geological Society Special Publication, **68**, 137-148.

Crame, J.A., Pirrie, D., Riding, J.B. & Thomson, M.R.A., 1991. Campanian-Maastrichtian (Cretaceous) stratigraphy of the James Ross Island area, Antarctica. *Journal of the Geological Society of London*, **148**, 1125-1140.

Cranwell, L.M., 1959. Fossil pollen from Seymour Island, Antarctica. *Nature*, **184**, 1782-1785.

Cranwell, L.M., Harrington, H.J. & Speden, I.G., 1960. Lower Tertiary microfossils from McMurdo Sound, Antarctica. *Nature*, **186**, 700-702.

Cranwell, L.M., 1964. Hystrichospheres as an aid to Antarctic dating with special reference to the recovery of *Cordosphaeridium* in erratics at McMurdo Sound. *Grana Palynologica*, **5 (3)**, 397-405.

Creber, G.T. & Chaloner, W.G., 1987. The contribution of growth rings to the reconstruction of past climates. *In*: Ward, R.G.W. (ed.). *Applications of tree-ring studies; current research in denrochronology and related subjects*. B.A.R. International Series 333, Oxford, 37-67.

Dale, B., 1983. Dinoflagellate resting cysts: "benthic plankton". *In*: Fryxell, G.A. (ed.). *Survival strategies of the Algae*. Cambridge University Press, Cambridge, England, 69-136.

Daly, M.C., Chorovizc, J. & Fairhead, J.D., 1989. Rift basin evolution in Africa: the influence of reactivated steep basement shear zones. *In*: Cooper, M.A. & Williams, G.F.D. (eds). *Inversion Tectonics*. Geological Society Special Publication, **44**, 309-334.

Daly, M.C., Lawrence, S.R., Kimuna, D. & Binga, M., 1991. Late Palaeozoic deformation in central Africa: a result of a distant collision? *Nature*, **350**, 605-607.

Davey, R.J. & Rogers, J., 1975. Palynomorph distribution in Recent offshore sediments along two traverses off South West Africa. *Marine Geology*, **18**, 213-255.

Davies, E.H., Bujak, J.P. & Williams, G.L., 1982. The application of dinoflagellates to paleoenvironmental problems. *Third North American Paleontological Convention, Proceedings*; **1**, 125-131.

Davis, J.C., 1986. *Statistics and Data Analysis in Geology*. Wiley, New York, pp. 646.

Dettmann, M.E., 1992. Structure and floristics of Cretaceous vegetation of southern Gondwana: implications for angiosperm biogeography. *Palaeobotanist*, **41**, 224-233.

Detmann, M.E., 1994. Cretaceous vegetation: the macrofossil record. *In*: Hill, R.S. (ed.). *History of the Australian vegetation, Cretaceous to Recent.* Cambridge University Press, 143-170.

Dettmann, M.E., Pocknall, D.T. Romero, E.J & del Carmen Zamaloa, M., 1990.

Nothofagidites Erdtman ex Potonié, 1960. A catalogue of species with notes on the paleogeographic distribution of *Nothofagus* Bl. (Southern Beech). *New Zealand Geological Survey Paleontological Bulletin*, **60**, 1-51.

Dingle, R.V. & Lavelle, M., 1998. Late Cretaceous-Cenozoic climatic variations of the northern Antarctic Peninsula: new geochemical evidence and review. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **141**, 215-232.

Dingle, R.V. & Lavelle, M., 2000. Antarctic Peninsula Late Cretaceous-Early Cenozoic palaeoenvironments and Gondwana palaeogeographics. *Journal of African Earth Sciences*, **31**, 91-105.

Doyle, P. & Whitham, A.G., 1991. Palaeoenvironments of the Nordenskjöld Formation: an Antarctic Late Jurassic-Early Cretaceous black shale-tuff sequence. *In*: Tyson, R.V. & Pearson, T.H. (eds). *Modern and ancient continental shelf anoxia*. Geological Society, London Special Publications, **58**, 397-414.

Elliot, D.H., 1988. Tectonic setting and evolution of the James Ross Basin, northern Antarctic Peninsula. *In* Feldmann, R.M. & Woodburne, M.O., (eds). *Geology and Paleontology of Seymour Island, Antarctic Peninsula*. Geological Society of America Memoir, **169**, 541-555.

Elliot, D.H. & Trautman, T.A., 1982. Lower Tertiary strata on Seymour Island, Antarctic Peninsula. *In* Craddock, C. (ed.). *Antarctic Geoscience*. Madison, Wisconsin, University of Wisconsin Press, 287-297.

Emery, D. & Myers, K.J., 1996. *Sequence Stratigraphy*. Blackwell Science, Oxford, pp. 297.

Farquharson, G.W., 1982. Late Mesozoic sedimentation in the northern Antarctic Peninsula and its relationship to the southern Andes. *Journal of the Geological Society of London*, **139**, 721-728.

Goodman, D.K., 1979. Dinoflagellate communities from the Lower Eocene of Nanjemoy Formation of Maryland, U.S.A. *Palynology*, **3**, 169-190.

Hall, S.A., 1977. Cretaceous and Tertiary dinoflagellates from Seymour Island, Antarctica. *Nature*, **267**, 239-241.

Harland, R., 1973. Dinoflagellate cysts and acritarchs from the Bearpaw Formation (Upper Campanian) of southern Alberta, Canada. *Palaeontology*, **16**, 665-706.

Harland, R., 1989. A dinoflagellate cyst record for the last 0.7 Ma from the Rockall Plateau, northeast Atlantic Ocean. *Journal of the Geological Society*, **146**, 945-951.

Harland, R., Pudsey, C.J., Howe, J.A. & Fitzpatrick, M.E.J., 1998. Recent dinoflagellate cysts in a transect from the Falkland Trough to the Weddell Sea, Antarctica. *Paleontology*. **41**, 1093-1131.

Hathway, B., 2000. Evolution of the Larsen Basin, Antarctica. *Journal of the Geological Society, London*, **157**, 417-432.

Helby, R., Morgan, R. & Partridge, A.D., 1987. A palynological zonation of the Australian Mesozoic. *Memoir of the Association of Australian Palaeontologists*, **4**, 1-94.

Hopkins, J.A. & McCarthy, F.M.G., 2000. Determining the resistance of fossil dinoflagellate cysts to oxidation: a laboratory approach. *Geological Society of America Abstracts*. No. 52828.

Hooper, P.R., Rehacek, J., Duncan, R.A., Marsh, J.S. & Duncan, A.R., 1993. The basalts of Lesotho, Karoo Province, Southern Africa. *EOS, Transactions of the American Geophysical Union*, **74**, 553.

Hudson, J.D., 1990. Salinity from faunal analysis and geochemistry. *In:* Briggs, D.E.G. & Crowther, P.R. (eds). *Palaeobiology: a synthesis*. Blackwell Scientific Publications, Oxford, 406-408.

Hyde, H.A., & Williams, D.A., 1944. The right word. *Pollen Analysis Circular*, **8**, 6.

Ineson, J.R., 1989. Coarse-grained submarine fan and slope apron deposits in a Cretaceous back-arc basin, Antarctica. *Sedimentology*, **36**, 793-819.

Ineson, J.R., Crame, J.A. & Thomson, M.R.A., 1986. Lithostratigraphy of the Cretaceous strata of west James Ross Island, Antarctica. *Cretaceous Research*, **7**, 141-159.

Keating, J.M., Spencer-Jones, M. & Newham, S., 1992. The stratigraphical palynology of the Kotick Point and the Whisky Bay formations, Gustav Group (Cretaceous), James Ross Island. *Antarctic Science*, **4**, 279-292.

Kidwell, S. M., 1989. Stratigraphic condensation of marine transgressive records: origin of major shell deposits in the Miocene of Maryland. *The Journal of Geology*, **97**, 1-24.

LaBreque, J.L., Cande, S., Bell, R., Raymond, C., Brozena, J., Keller, M., Parra, J.C. & Yanez, G., 1986. Aerogeophysical survey yields new data in the Weddell Sea. *Antarctic Journal*, **21**, 69-70.

Lawver, L.A. & Gahagan, L.M., 1998. Opening of Drake Passage and its impact on Cenozoic Ocean Circulation. *In*: Crowley, T.J. & Burke K.C. (eds*). Tectonic Boundary Conditions for Climatic Reconstructions.* Oxford University Press, Oxford Monographs on Geology and Geophysics, **39**, 212-223.

Macdonald, D.I.M. & Butterworth, P.J., 1990. The stratigraphy, setting and hydrocarbon potential of the Mesozoic sedimentary basins of the Antarctic Peninsula. *In*: St.John, B. (ed.). *Antarctica as an Exploration Frontier.* American Association of Petroleum Geologists, Studies in Geology, **139**, 100-105.

Macellari, C.E., 1988. Stratigraphy, sedimentology and palaeoecology of Upper Cretaceous/Paleocene shelf-deltaic sediments of Seymour Island. *In:* Feldmann, R.M. & Woodburne, M.O., (eds). *Geology and Paleontology of Seymour Island, Antarctic Peninsula.* Geological Society of America Memoir, **169**, 25-53.

MacPhail, M.K., Alley, N.F., Truswell, E.M., Sluiter, I.R.K., 1994. Early Tertiary vegetation: evidence from spores and pollen. *In*: Hill, R.S. (ed.). *History of the Australian vegetation.* Cambridge University Press, 189-261.

Manum, S.B., Boulter, M.C., Gunnarsdottir, H., Ranges, K. & Scholze, A., 1989. 32. Eocene to Miocene palynology of the Norwegian Sea (ODP Leg 104). *In*: Eldholm, O., Thiede, J., Taylor, E. *et al.* (eds). *Ocean Drilling Program, Proceedings, Scientific Results*, **104**, 611-639.

Marenssi, S.A., Santillana, S.N. & Rinaldi, C.A., 1998. Paleoambientes sedimentarios de la Aloformacion La Meseta (Eoceno), Isla Marambio (Seymour), Antartida. *Contribución Cientifica del Instituto Antártico Argentino*, **464**, 1-51.

Marret, F. & De Vernal, A., 1997. Dinoflagellate cyst distribution in surface sediments of the southern Indian Ocean. *Marine Micropaleontology*, **29**, 367-392.

Mildenhall, D.C., 1989. Terrestrial Palynology. *In*: Barrett, P.J. (ed.). Antarctic Cenozoic history from the CIROS-1 drillhole, McMurdo Sound. *DSIR Bulletin* **245**, 119-127.

Molnar, R.E., 1989. Terrestrial tetrapods in Cretaceous Antarctica. *Geological Society Special Publication*, **47**, 131-140.

Mudie, P.J., 1986. Palynology and dinoflagellate biostratigraphy of Deep Sea Drilling Project Leg 94, Sites 607 and 611, North Atlantic Ocean. *In*: Ruddiman, W.F., Kidd, R.B., Thomas, E. *et al.*(eds). *Deep Sea Drilling Project, Initial Reports*, **94**, 785-812.

Nederbragt, A.J., Fiorentino, A. & Klosowska, B., 2001. Quantitative analysis of calcareous microfossils across the Albian-Cenomanian boundary oceanic anoxic event at DSDP Site 547 (North Atlantic). *Palaeogeography, Palaeoclimatology, Palaeoecology*, **166**, 401-421.

Nix, H., 1982. Environmental determinants of biogeography and evolution in Terra Australia. *In*: Barker, W.R & Greenslade, P.J.M. (eds). *Evolution of the flora and fauna of arid Australia.* Frewville, Peacock Publications, 47-66.

Palamarczuk, S., 1982. Dinoflagelados de edad Daniana en la isla Vicecomodoro Maramabio (ex-Seymour), Antártida Argentina. *Ameghiniana*, **19**, 353-360.

Palamarczuk, S., Ambrosini, G., Medina, F.A., Martinez Macchiavello, J.C. & Rinaldi C.A., 1984. Las formaciones López de Bertodano y Sobral en la isla Vicecomodoro Marambio, Antártida. *Noveno Congreso Geológico Argentino Actas*, **1**, 399-419.

Pankhurst, R.J., 1982. Rb-Sr geochronology of Graham Land, Antarctica. *Journal of the Geological Society, London*, **139**, 701-711.

Pirrie, D., Witham, A.G. & Ineson, J.R., 1991. The role of tectonics and eustacy in the evolution of a marginal basin: Cretaceous-Tertiary Larsen Basin, Antarctica. *In*: MacDonald, D.I.M. (ed.), Sedimentation, Tectonics and Eustacy: Sea-level changes at Active Margins. *International Association of Sedimentologists, Special Publications*, **12**, 293-305.

Pirrie, D. Duane, A.M. & Riding, J.B., 1992. Jurassic-Tertiary stratigraphy and palynology of the James Ross Basin: review and introduction. *Antarctic Science*, **4**, 259-266.

Pocknall, D.T., 1989. Late Eocene to Early Miocene vegetation and climate history of New Zealand. *Journal of the Royal Society of New Zealand*, **19**, 1-18.

Porebski, S.J., 1995. Facies architecture in a tectonically-controlled incised valley estuary: La Meseta Formation (Eocene) of Seymour Island, Antarctic Peninsula. *Studia Geologica Polonica*, **107**, 7-97.

Porebski, S.J., 2000. Shelf-valley compound fill produced by fault subsidence and eustatic sea-level changes, Eocene La Meseta Formation, Seymour Island, Antarctica. *Geology*, **28**, 147-150.

Powell, A.J., Lewis, J. & Dodge, J.D., 1992. The palynological expressions of post-Paleogene upwelling: a review. *In*: Summerhayes, C.P., Prell, W.L. & Emeis, K.C. (eds). Upwelling systems: evolution since the Early Miocene. *Geological Society of London, Special Publication*, **64**, 215-226.

Prauss, M., 1989. Dinozysten-Stratigraphie und Palynofazies im oberen Lias und Dogger von NW-Deutschland. *Palaeontographica, Abteilung B*, **218**, 93-191.

Rapela, C.W., & Pamkhurst, R.J., 1992. The granites of northern Patagonia and the Gastre Fault System in relation to the break-up of Gondwana. *In:* Storey, B.C.,

Alabaster, T. & Pankhurst, R. (eds). *Magmatism and the Causes of Continental Break-up*. Geological Society Special Publication, **68**, 209-220.

Rich, T.H., Rich, P.V., Wagstaff, B., McEwen-Mason, J., Douthitt, C.B. & Gregory, R.T., 1989. Early Cretaceous biota from the northern side of the Australo-Antarctic rift valley. *Geological Society Special Publication*, **47**, 121-130.

Riding, J. B., Crame, J. A., Dettmann, M. E. & Cantrill, D.J., 1998. The age of the base of the Gustav Group in the James Ross Basin, Antarctica. *Cretaceous Research*, **19**, 87-105.

Rinaldi, C.A., Massabie, A., Morelli, J., Rosenmall, H.L. & del Valle, L., 1978. Geologia de la Isla Vicecomodoro Marambio. *Instituto Antático Argentino, Contribuciones*, **217**, 1-37.

Roeser, H.A., Fritsch, J. & Hinz, K., 1996. The development of the crust off Dronning Maud Land, East Antarctica. *In:* Storey, B.C., King, E.C. & Livermore, R.A. (eds). *Weddell Sea Tectonics and Gondwana Break-up*. Geological Society Special Publication, **108**, 243-264.

Sadler, P.M., 1988. Geometry and stratification of uppermost Cretaceous and Paleogene units on Seymour Island, northern Antarctic Peninsula. *In:* Feldmann, R.M. & Woodburne, M.O., (eds). *Geology and Paleontology of Seymour Island, Antarctic Peninsula*. Geological Society of America Memoir, **169**, 303-320.

Shackleton, N.J. & Kennett, J.P., 1975. Paleotemperature history of the Cenozoic and the initiation of Antarctic glaciation, oxygen and carbon isotope analyses in DSDP sites 277, 279 and 281. *Initial Reports of the Deep Sea Drilling Project*, **29**. 743-755.

Simpson, E.S.W., Sclater, J.G., Parsons, B., Nortin, I.O. & Meinke, L., 1979. Mesozoic magnetic lineations in the Mozambique Basin. *Earth and Planetary Science Letters*, **43**, 260-264.

Snape, M.G., 1992. Dinoflagellate cysts from an allochthonous block of Nordenskjöld Formation (Upper Jurassic), north-west James Ross Island. *Antarctic Science*, **4**, 337-353.

Specht, R.I., Dettmann, M.E. & Jarzen, D.M, 1992. Community associations and structure in the Late Cretaceous vegetation of southeast Australasia and Antarctica. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **94**, 283-309.

Stockmarr, J., 1971. Tablets with spores used in absolute pollen analysis. *Pollen et Spores*, **13**, 615-621.

Storey, B.C. & Garrett, S.W., 1985. Crustal growth of the Antarctic Peninsula by accretion, magmatism and extension. *Geological Magazine*, **122**, 5-14.

Storey, B.C, Alabaster, T., Hole, M.J., Pamkhurst, R.J. & Wever, H.E., 1992. Role of subduction-plate boundary forces during the initial stages of Gondwana break-up: evidence from the proto-Pacific margin of Antarctica. *In:* Storey, B.C., Alabaster, T. & Pankhurst, R. (eds). *Magmatism and the Causes of Continental Break-up*. Geological Society Special Publication, **68**, 149-163.

Storey, B.C., Vaughan, A.P.M., & Millar, I.L., 1996. Geodynamic evolution of the Antarctic Peninsula during Mesozoic times and its bearing on Weddell Sea history. *In:* Storey, B.C., King, E.C & Livermore, R.A. (eds). *Weddell Sea Tectonics and Gondwana Break-Up*. Geological Society Special Publication, **108**, 87-103.

Storey, B.C., Leat, P.T., Weaver, S.D., Pankhurst, R.J., Bradshaw, J.D. & Kelley, S., 1999. Mantle plumes and Antarctica-New Zealand rifting: evidence from mid-

Cretaceous mafic dykes. *Journal of the Geological Society, London*, **156**, 659-671.

Stover, L.E., Brinkhuis, H., Damassa, S.P., de Verteuil, L., Helby, R.J., Monteil, E. Partridge, A.D., Powell, A.J., Riding, J.B., Smelror, M. & Williams, G.L., 1996. Mesozoic-Tertiary dinoflagellates, acritarchs and prasinophytes. *In:* Jansonius, J. & McGregor, D.C. (eds). *Palynology: principles and applications.* American Association of Stratigraphic Palynologists Foundation, **2**, 641-750.

Strother, P.K., 1996. Chapter 5. Acritarchs. *In*: Jansonius, J. & McGregor, D.C. (eds). *Palynology: principles and applications.* American Association of Stratigraphic Palynologists Foundation, **1**, 81-106.

Taylor, F.J.R., 1990. Phylum Dinoflagellata. *In*: Margulis, L., Corliss, J.O., Melkonian, M., Chapman, D.J. & MckHann, H.I. (eds). *Handbook of Protoctista.* Jones & Bartlett, Boston, 419-437.

Thomson, M.R.A. & Farquharson, G.W., 1984. Discovery and significance of the ammonite genus *Favrella* in the Antarctic Peninsula area. *British Antarctic Survey Bulletin*, **62**, 7-14.

Traverse, A., 1988. *Paleopalynology.* Chapman and Hall, pp 600.

Truswell, E. M., 1986. Palynology. *In*: Barrett, P. J. (ed.) Antarctic Cenozoic history, MSSTS-1 drillhole. *DSIR bulletin*, **237**, 131-134.

Tyson, R.V., 1995. *Sedimentary organic matter: organic facies and palynofacies.* Chapman and Hall, pp 615.

Versteegh, G.J.M., 1994. Recognition of cyclic and non-cyclic environmental changes in the Mediterranean Pliocene: A palynological approach. *Marine Micropaleontology*, **23**, 147-183.

von Post, L., 1917. Om skogsträdpollen i sydsvenska torFormationosselagerföljder. *Geologiska Föreningens i Stockholm Förhandlingar*, **38**, 384-390.

Wall, D., 1965. Microplankton, pollen, and spores from the Lower Jurassic in Britain. *Micropaleontology*, **11**, 151-190.

Wall, D., Dale, B., Lohmann, G.P. & Smith, W.K., 1977. The environmental and climatic distribution of dinoflagellate cysts in modern marine sediments from regions in the North and South Atlantic Oceans and adjacent areas. *Marine Micropaleontology*, **2**, 121-200.

Wilson, G.J., 1989. Marine Palynology. *In:* Barrett, P.J. (ed.). Antarctic Cenozoic history from the CIROS-1 drillhole, McMurdo Sound. *DSIR Bulletin*, **245**, 129-133.

Wilson, M., 1992. Magmatism and continental rifting during the opening of the South Atlantic Ocean: a consequence of Lower Cretaceous super-plume activity? *In:* Storey, B.C., Alabaster, T. & Pankhurst, R. (eds). *Magmatism and the Causes of Continental Break-up.* Geological Society Special Publication, **68**, 241-255.

Whitham, A.G. & Doyle, P., 1989. Stratigraphy of the Upper Jurassic-Lower Cretaceous Nordenskjöld Formation of eastern Graham Land, Antarctica. *Journal of South American Earth Sciences*, **2**, 371-384.

Willumsen, P.S., 2000. Late Cretaceous to Early Paleocene palynological changes in midlatitude Southern Hemisphere, New Zealand. *GFF*, **12**, 180-181.

Wrenn, J.H., & Hart, G.F., 1988. Paleogene dinoflagellate cyst biostratigraphy of Seymour Island, Antarctica. *In:* Feldmann, R.M. & Woodburne, M.O., (eds). *Geology and Paleontology of Seymour Island, Antarctic Peninsula.* Geological Society of America Memoir, **169**, 321-447.

Vaughan, A.P.M. & Storey, B.C., 2000. The eastern Palmer Land shear zone: a new terrane model for the Mesozoic development of the Antarctic Peninsula. *Journal of the Geological Society, London*, **157**, 1243-1256.

Zinsmeister, W.J., 1988. Early geological exploration of Seymour Isalnd, Antarctica. *In*: Feldmann, R.M. & Woodburne, M.O., (eds). *Geology and Paleontology of Seymour Island, Antarctic Peninsula*. Geological Society of America Memoir, **169**, 1-16.

Zachos, J.C., Stott, L.D., Lohmann, K.C., 1994. Evolution of early Cenozoic marine temperatures. *Paleoceanography*, **9 (2)**, 353-387.

THESIS CONTAINS CD/DVD